Arctic Ocean Circulation☆

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Background

The Arctic Ocean, the smallest of the world oceans, was until recently the one most unknown. The scant existing knowledge was gained from shattered attempts to find and navigate the Northeast and Northwest passages and from frustrated expeditions to reach the North Pole. Ice stopped them all, and the Arctic Ocean remained a white, likely ice covered, spot on the world map.

Strangely enough, in the 19th century it was seriously discussed that the Arctic Ocean, beyond the ice encountered by the expeditions, might be an open ocean. One argument for this view was that the ocean currents, especially in the North Atlantic but also in the North Pacific, could bring enough heat to the Arctic Ocean to melt the ice beyond the runoff dominated areas close to the continents. The expeditions searching this open sea, either through Barents Sea or between Greenland and Svalbard, beyond Baffin Bay and through Bering Strait, all failed. It was not until 1893–96, when Fridtjof Nansen let *Fram* be frozen in and drift with the ice from the Laptev Sea to the passage between Greenland and Svalbard, later named Fram Strait, that the idea, or dream, of an open Arctic Ocean was finally dispelled. However, Nansen observed a warm subsurface layer beneath the 100 m thick, cold and low salinity surface water. Warm Atlantic water was present in the Arctic Ocean but covered by low salinity water, deriving from runoff and/or from sea ice melting, which prevented the heat from the Atlantic layer to reach the sea surface and the ice. This observation focuses one of the most fundamental questions in Arctic oceanography: Why does not the heat present in the Atlantic layer melt the ice, and could the conditions in the Arctic change so that the Atlantic heat becomes available to remove the ice cover? This question is raised again today in connection with the observed diminished sea ice extent.

The Fram expedition contributed tremendously to the knowledge of the high latitude Arctic Ocean but apart from the accidental drift of the Soviet ice breaker Sedov subsequent studies were confined to the boundaries of the Arctic Ocean, to the Norwegian and Greenland seas, the Barents Sea, and along the coasts of Siberian continent and the Canadian Arctic Archipelago. *Vega* had sailed the Northeast passage already 1878–80 but not until 1903–1905 did *Gjøa* navigate the Northwest passage. However, several aircrafts missions to the high Arctic Ocean were conducted, mostly by the Soviet Union, and in 1937 a crucial step was taken with the establishment of the first Soviet drifting ice station, North Pole 1, at the North Pole. This station was followed by several others, mainly launched by the Soviet Union but also by the United States and Canada, and most of the exploration of the Arctic Ocean between 1937 and the 1980s was made from ice stations and from aircraft missions, conducted in spring. In 1980, one of the first scientific icebreaker expeditions, YMER-80, was organized to celebrate the 100 years anniversary of *Vega's* voyage. It was followed by several others and in 1991 the research icebreakers *Oden* and *Polarstern* jointly reached the North Pole. Icebreakers are free to navigate in the ice, and the knowledge of water masses and circulation as well as of the bathymetry increased substantially. Scientific submarine expeditions also began in the 1980s, adding information of hydrography and sea ice thickness.

The 1980s and 1990s saw the beginning of large international research programs, first in the marginal ice zone but later in the high Arctic. These efforts culminated in the fourth International Polar Year (IPY) 2007–2009. During this period observational techniques and instrumentation developed rapidly. More parameters were measured and the data coverage increased enormously.

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Especially the advent of satellites and the use of ice tethered platforms contribute almost online data gathering from the Arctic Ocean. These observations, especially of the growth and decline of the sea ice cover, have made the general public acutely aware of the state and the ongoing changes in the Arctic Ocean.

Geography and Bathymetry and the Connections With the World Ocean

The Arctic Ocean is the northernmost part of the North Atlantic. Its boundary to the south comprises the northern coast of Greenland, Fram Strait, the west coast of Svalbard and the shelf break between Svalbard and Norway, the Eurasian coastline, Bering Strait, the North American coastline and the Canadian Arctic Archipelago. About 53% of its area of 9.5×10^{12} m^2 consists of shallow shelf seas, Barents Sea (mean depth 200 m), Kara Sea (130 m), Laptev Sea (50 m), East Siberian Sea (60 m), Chukchi Sea (80 m) and the North American shelf. The remaining 47% are deep basins. The two major basins, the Eurasian Basin and the Amerasian Basin, are separated by the Lomonosov Ridge with sill depth 1870 m. The Eurasian Basin consists of two basins, the about 4000 m deep Nansen Basin and the 4500 m deep Amundsen Basin separated by the Gakkel Ridge, which includes the deepest trench of the Arctic Ocean, more than 5000 m deep. The Amerasian Basin also has two major basins, the large Canada Basin, 3800 m deep and the smaller, 4000 m deep Makarov Basin. They are connected through a 2400 m deep gap between the Alpha Ridge and the Mendeleyev Ridge (Fig. 1).

The main communication between the Arctic Ocean and the world ocean is through the 2600 m deep Fram Strait and over the 250 m deep Barents Sea, which connect the Eurasian Basin to the Nordic Seas (The Greenland, Iceland and Norwegian seas), separated from the North Atlantic proper by the Greenland-Scotland Ridge with sill depth 500 to 840 m. There is a second passage to the North Atlantic west of Greenland, through the shallow and narrow Nares Strait (sill depth 240 m), Jones Sound (150 m) and Lancaster Sound (135 m) that connect the Amerasian Basin to Baffin Bay and then across the 640 m deep sill in Davis Strait to the

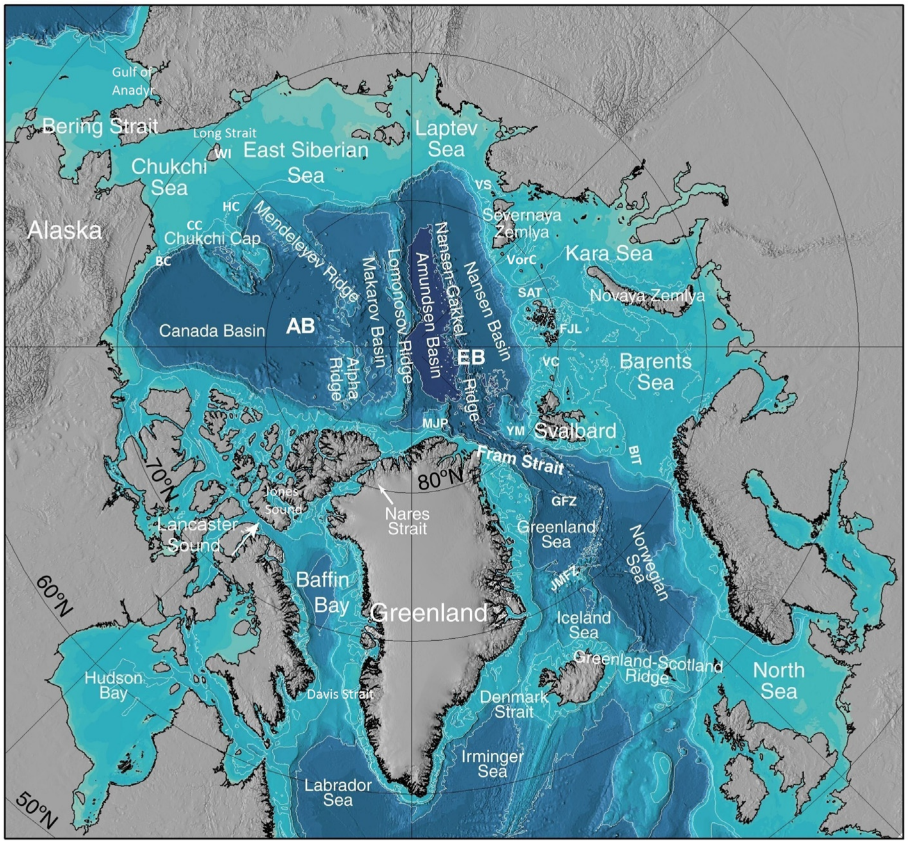

Fig. 1 Map of the Arctic Mediterranean Sea showing geographical and bathymetric features. The bathymetry is from the IBCAO updated data base and the projection is Lambert Equal Area. The 200, 500, 2000, and 4000 m isobaths are shown. *AB*: Amerasian Basin; *BIT*: Bear Island Trough (Barents Sea opening); *BC*: Barrow Canyon; *CC*: Central Channel; *EB*: Eurasian Basin; *FJL*: Franz Josef Land; *GFZ*: Greenland Fracture Zone; *HC*: Herald Canyon; *MJP*: Morris Jessup Plateau, *JMFZ*: Jan Mayen Fracture Zone; *SAT*: St. Anna Trough; *YP*: Yermak Plateau; VC: Victoria Channel; *VorC*: Voronin Canyon; *VS*: Vilkitskij Strait; *WI*: Wrangel Island. Modified from Rudels, B., Anderson, L. G. and Eriksson, P. et al. (2012). Observations in the ocean. In: Lemke, P. & Jacobi, H.-W. (eds.) *Arctic climate change—The ACSYS decade and beyond*, pp. 117–198. Heidelberg: Springer.

Labrador Sea. There is a passage from the North Pacific to the Arctic Ocean through the narrow (85 km) and shallow (45 m) Bering Strait, connecting two extremes of the world ocean, the convectively active and ventilated North Atlantic and the stratified, stagnant North Pacific.

The waters in the Arctic Ocean mainly derive from the North Atlantic, crossing the Greenland-Scotland Ridge and then flowing through the Norwegian Sea in the Norwegian Atlantic Current. As the Norwegian Atlantic Current reaches the Barents Sea opening between Norway and Bear Island it splits. One part enters, together with the Norwegian Coastal Current, the Barents Sea, while the rest continues as the West Spitsbergen Current to Fram Strait. The smaller volume of Pacific water entering through Bering Strait provides less saline and less dense waters that mainly occupy the upper part of the Canada Basin. Pacific derived waters can be further identified by additional tracers. Pacific water has higher Silicate content than Atlantic water and its Nitrate/Phosphate ratio is lower, because Nitrate is reduced in the Oxygen poor North Pacific. In addition, freshwater is provided by the atmosphere, mainly as river runoff, draining a catchment area larger than the Arctic Ocean, but partly as direct precipitation. The freshwater input creates the strong stability that allows sea ice to form in winter and be maintained throughout the year.

The outflows to the world ocean occur in the East Greenland Current in the western Fram Strait, which carries most of the ice export and about half of the liquid freshwater export and almost all of the returning but transformed Atlantic water, and occasionally also Pacific water. The East Greenland Current continues along the east coast of Greenland until Cape Farvel, where it turns north as the West Greenland Current and partly enters Baffin Bay to eventually join the low salinity, mainly Pacific, waters that have passed through the Canadian Arctic Archipelago west of Greenland. All these waters then flow south in the Baffin Island Current crossing Davis Strait into the Labrador Sea (Fig. 2).

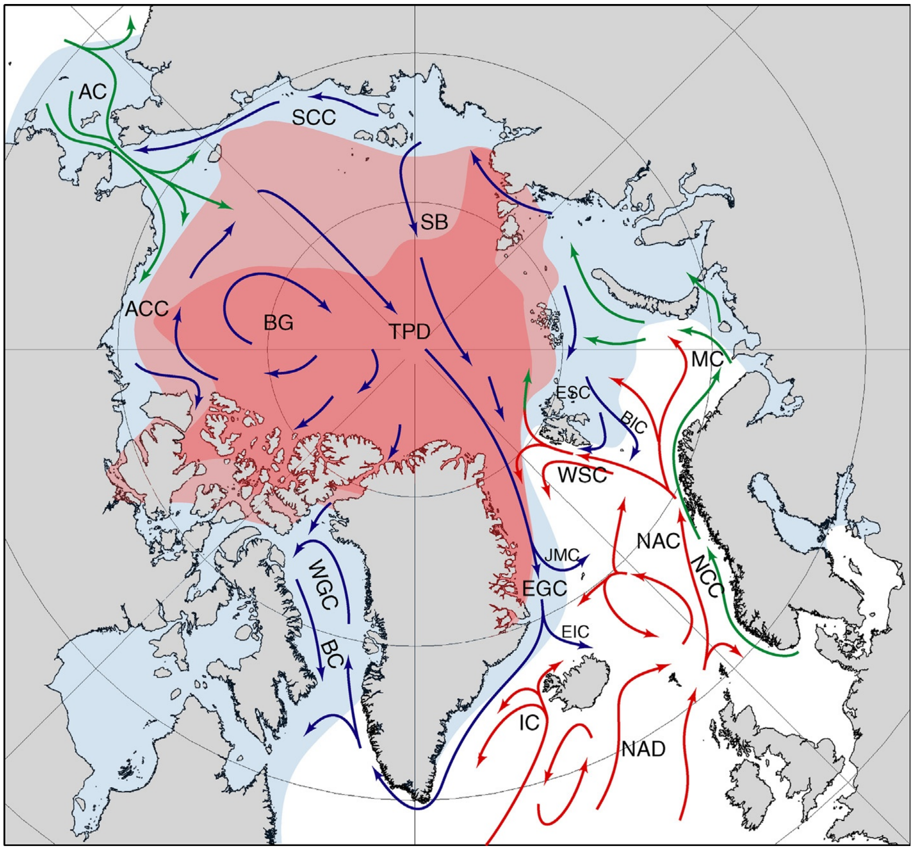

Fig. 2 The circulation of the upper layers of the Arctic Mediterranean Sea. Warm Atlantic currents are indicated by *red arrows*, cold less saline polar and arctic currents by *blue arrows*. Low salinity transformed currents are shown by *green arrows*. The maximum ice extent is shown in *blue* and the minimum ice extent in *red* (late 20th century conditions). The minimum in 2007, the second absolute minimum to date, is shown in *dark red*. *AC*: Anadyr Current; *ACC*: Alaskan Coastal Current; *BC*: Baffin Island Current; *BIC*: Bear Island Current; *BG*: Beaufort Gyre; *EGS*: East Greenland Current; *EIC*: East Iceland Current; *ESC*: East Spitsbergen Current; *IC*: Irminger Current; *JMC*: Jan Mayen Current; *MC*: Murman Current; *NAD*: North Atlantic Drift; *NAC*: Norwegian Atlantic Current; *NCC*: Norwegian Coastal Current; *SB*: Siberian branch (of the Transpolar Drift); *SCC*: Siberian Coastal Current; *TPD*: Transpolar Drift; *WGC*: West Greenland Current; *WSC*: West Spitsbergen Current. From Rudels, B., Anderson, L. G. and Eriksson, P. et al. (2012). Observations in the ocean. In: Lemke, P. & Jacobi, H.-W. (eds.) *Arctic climate change—The ACSYS decade and beyond*, pp. 117–198. Heidelberg: Springer.

Processes

The most characteristic and most important feature of the Arctic Ocean is its perennial ice cover. The Arctic is dominated by net heat loss to space, leading to harsh climate with strong seasonal variations, alternating between the cold, dark polar night in winter and the light, cool summer day. Furthermore, the input of freshwater, by the very seasonal (summer) river runoff and by net precipitation, creates a less saline and shallow surface water, and the stratification is strong enough to allow the upper layer to cool to freezing temperature in winter without convection into the deep. Instead sea ice forms and grows throughout the winter. The heat stored in the deeper layers becomes isolated and the heat provided to the atmosphere, and to space, is latent heat of freezing. In summer the ice cover reflects a substantial fraction of the incoming solar radiation. The absorbed heat is mainly used to melt ice, creating a low salinity, upper melt water layer, where the temperature is kept close to freezing. Although the radiation balance in summer is positive (the Arctic Ocean is gaining heat) both the outgoing and the back radiated long wave radiation are larger than the incoming and reflected solar radiation. The positive radiation balance is used, by melting sea ice, to "pay back" the latent heat released by ice formation in winter. The annual freezing and melting of sea ice is about 1.5 m, and since the ice cover remains throughout the year and sea ice is exported to lower latitudes, ice formation in winter is a net heat source to the Arctic.

Most of the Arctic Ocean is dominated by net ice formation. However, the Nansen Basin is an exception. In the western Nansen Basin, north of Svalbard, the Atlantic inflow encounters, and melts, sea ice, and its upper part becomes transformed into a less saline, colder layer, the polar mixed layer (PML). This occurs primarily in winter and the heat lost by the Atlantic water is partly used to melt sea ice and partly lost to atmosphere and space. The fraction going to ice melt depends on, and increases with, the temperature of the Atlantic water. This interaction between Atlantic water, sea ice and atmosphere results in the largest Atlantic water heat loss in the deep Arctic basins. In the following summer seasonal ice melt creates a low salinity melt water layer that is removed in fall by ice formation and the upper layer is homogenized by brine rejection down to the thermocline, which in the Nansen Basin separates the Atlantic water from the PML. As long as the thermocline and halocline coincide, heat can be transferred from the Atlantic water to the PML in winter and be brought to the surface by the stirring generated by the haline convection.

The thermocline in the Nansen Basin is characterized by thin "diffusive" interfaces with large temperature, ΔT, and salinity, ΔS, steps, small density ratios, $R_\rho = \beta\Delta S/\alpha\Delta T$ (α and β being the coefficients of heat expansion and salinity contraction, respectively), separated by thick homogenous layers, suggesting upward heat transports by double diffusive convection. Diffusive interfaces are also present in the Amerasian Basin but here the layers are thinner, the interfaces thicker, the temperature steps smaller and the density ratios larger. Possible diffusive transports are there expected to be much weaker than in the Nansen Basin.

North of the Laptev Sea less saline shelf water crosses the shelf break and covers the PML advected with the Atlantic water eastward along the continental slope. The haline convection in winter homogenizes and transforms the shelf water to the PML present in the Amundsen and Makarov basins, but the convection does not reach the PML advected from the Nansen Basin, which becomes a subsurface, cold halocline water mass. Beyond the Nansen Basin the thermocline and halocline no longer coincide and heat transferred upward from the Atlantic layer is trapped in the halocline layer (Fig. 3).

The Atlantic water at the continental slope north and east of the Kara Sea is characterized by strong interleaving, not only in the boundary current but also in the interior of the basins. The intrusions in the boundary current are created at the confluence of the two Atlantic branches north of Severnaya Zemlya (see circulation below), but their coherence and extension across the Eurasian Basin are subject to different explanations (Fig. 4). Theoretical work has shown that intrusions can be created by double diffusive convection, if one components, heat or salt, is unstably stratified. However, in the Arctic Ocean intrusions are found in all possible background stratifications, not only at diffusive and/or salt finger unstable but also at stable–stable stratification. In the stable–stable situation sharp fronts and large lateral displacements might be necessary to create the initial interleaving. Once intrusions are formed, alternating diffusive and salt finger interfaces transport buoyancy vertically between the layers, generating lateral pressure gradients that drive the intrusions across the front. The unanswered question is; how important are these intrusions in spreading heat from the boundary current into the basin interior? Are they self-propelled and reach throughout the basin, or is the double-diffusively driven motion only active over a short period and distance until the potential energy stored in the unstably distributed component has been removed, after which the interleaving structures become fossil and are advected by the mean circulation in the interior of the basins (Fig. 4)?

The haline convection is in the basins limited to the PML. However, on the shelves ice drift caused by local winds can create open water, lee polynyas, close to islands and coasts, as well as flaw polynyas between fast ice and drifting ice. In both situations more ice is formed and more brine is rejected. The haline convection reaches the bottom, increasing the salinity of the underlying water, and eventually brine enriched shelf water crosses the shelf break and sinks into the deep basins as entraining boundary plumes. The depth reached by the plumes depends upon the amount of released brine, the depth of the shelf and the amount of ambient water entrained during the descent. Most of the dense shelf water supplies the halocline layers but some penetrates down to and also beneath the Atlantic water, cooling the Atlantic layer and redistributing warm Atlantic water downward to deeper layers.

The observed downstream reduction of the Atlantic water temperature beyond the Gakkel Ridge is then not primarily caused by an upward heat flux but rather due to the mixing between the two inflow branches and by the incorporation of cold shelf water plumes. The lower temperature is balanced by a larger volume and the heat stays in the Atlantic layer and is largely conserved.

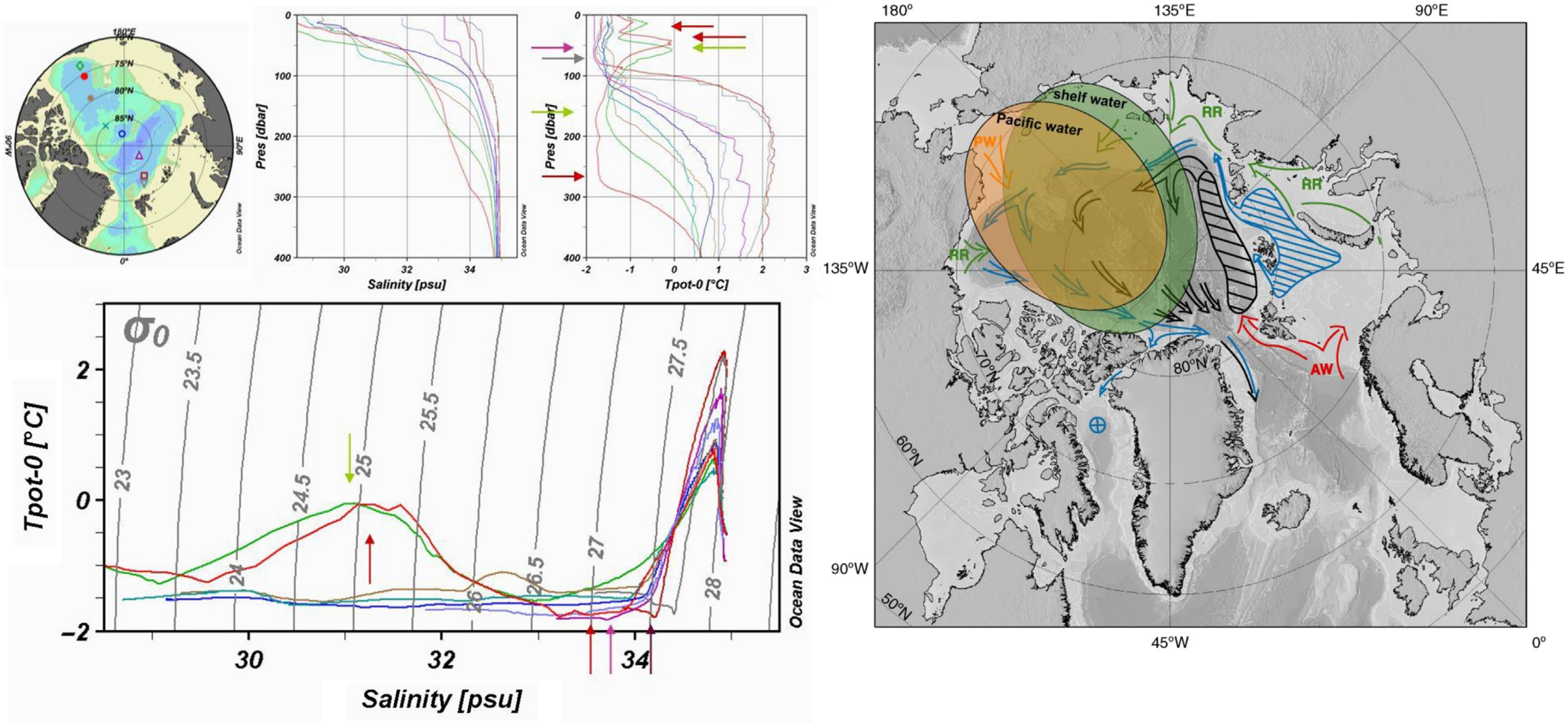

Fig. 3 Left panel: Profiles and TS curves showing halocline features in the Arctic Ocean. The *gray*, *dark red* and *magenta arrows* indicate the temperature minimum created by winter homogenization. The *red arrows* show on one station the upper temperature maximum due to seasonal heating, the near surface temperature maximum (NSTM), the temperature maximum due to the inflow of Bering Seat Summer Water (BSSW) and a temperature minimum due to the presence of an (isolated?) eddy of Fram Strait branch lower halocline water. The *green arrows* show the temperature maximum of the BSSW and the temperature minimum of the Bering Sea Winter Water (BSWW). Note that the NSTM is present also at the green station and that the temperature minimum of the winter homogenization is present above the BSSW maximum on the *green* and *red stations*. Right panel: The sources and circulation of halocline waters in the Arctic Ocean. The schematic shows the formation areas and the circulation of the waters created by sea ice melting in the Nansen Basin and northern Barents Sea, and how they are covered by less saline shelf water and Pacific water to become the Fram Strait branch and the Barents Sea branch lower halocline waters. Modified from Rudels, B. (2012). Arctic Ocean circulation and variability—Advection and external forcing encounter constraints and local processes. *Ocean Science* **8**, 261–286. https://doi.org/10.5194/os-8-261-2012.

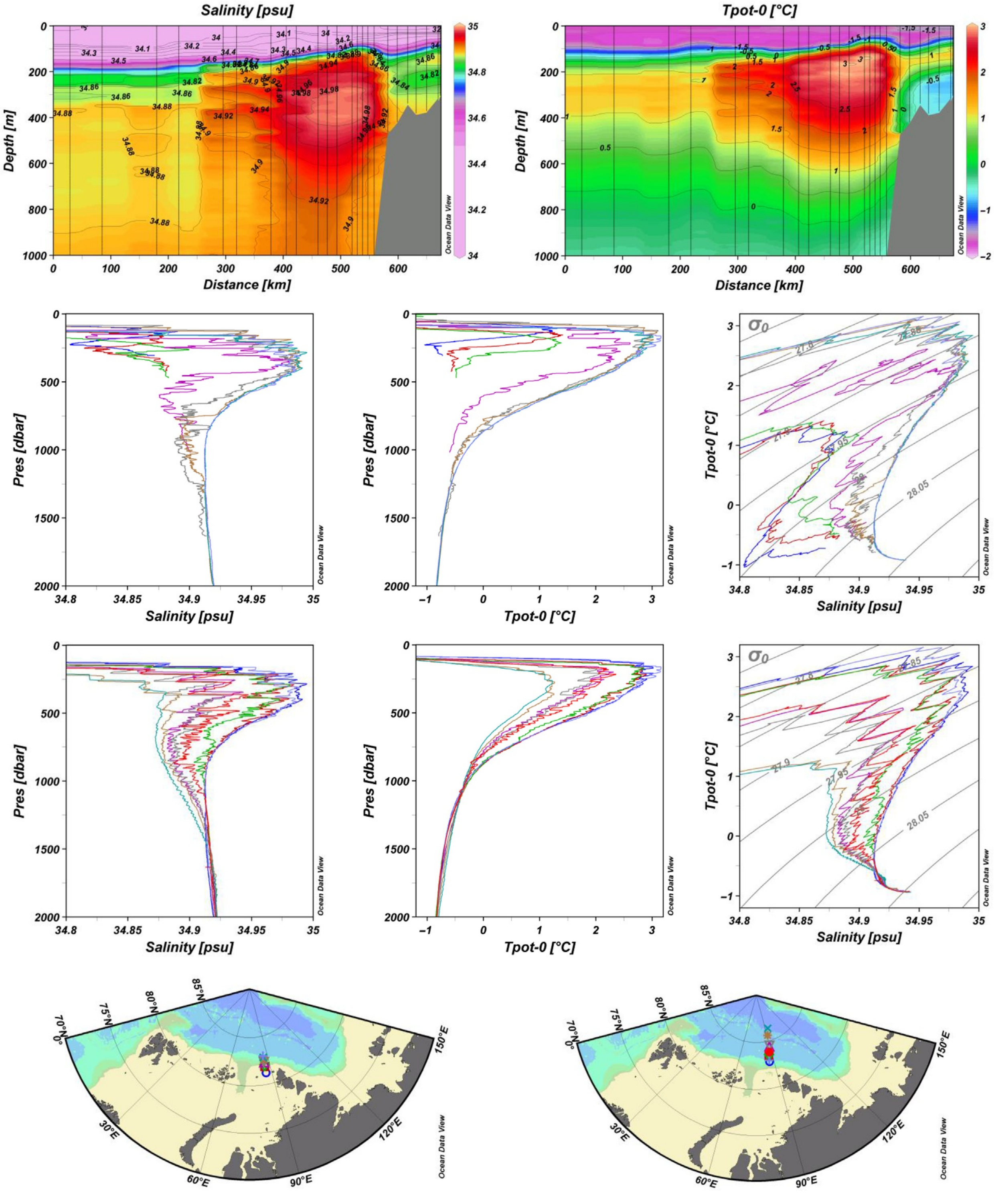

Fig. 4 Water masses, mixing and interleaving in the Nansen Basin north of the eastern Kara Sea. Upper panel: Temperature and salinity sections from the shelf break to the Gakkel Ridge showing the warm, saline Fram Strait Atlantic core and the cooler and less saline Barents Sea branch water on the slope but also on the basin side of the warm core. The interleaving structures on the basin side of the salinity and temperature maxima are assumed advected with the mean circulation towards Fram Strait. Upper central panel: Potential temperature and salinity profiles and TS curves from the slope showing the characteristics of the two branches and the mixing between them across the front. The less saline layer between 500 and 1500 m below the Fram Strait salinity maximum derives from the denser Barents Sea branch water entering the Nansen Basin via the St. Anna Trough. Lower central panel: Potential temperature and salinity profiles and TS curves from the deep Nansen and Amundsen basins showing the interleaving structures and different fronts. Lower panel, positions of the stations on upper central panel (left) and lower central panel (right). Modified from Rudels, B. (2012). Arctic Ocean circulation and variability—Advection and external forcing encounter constraints and local processes. *Ocean Science* **8**, 261–286. https://doi.org/10.5194/os-8-261-2012.

Circulation

The waters entering the Arctic Ocean have their own characteristic circulation paths and transformation history, which have to be considered separately.

The Upper Layers

The movements of the ice cover and the upper layers of the Arctic Ocean are mainly driven by the wind. The atmospheric circulation over the Amerasian Basin is predominantly anti-cyclonic with the high pressure cell centered over the Canada Basin. The wind forces the ice and the underlying water to the right of the wind, a phenomenon first observed by Fridtjof Nansen on *Fram* and later explained as an effect of the rotation of the earth by Vagn Walfrid Ekman. This creates a convergence and accumulation of upper layer water in the Canada Basin, the Beaufort Gyre, and the resulting pressure field and geostrophic circulation extends at least down to 300–400 m depth. The ice eventually escapes from the gyre in the Transpolar Drift and exits through Fram Strait. The atmospheric circulation over the Eurasian Basin, by contrast, is influenced by low pressure systems moving from the North Atlantic through the Nordic Seas to the Arctic Ocean. The circulation in the Eurasian Basin is mostly cyclonic and adds sea ice and upper layer water from the Siberian shelves to the Transpolar Drift, the Siberian branch (Fig. 2).

The shelf water has two components, the river runoff and a saline component supplied either from the North Atlantic by the Norwegian Coastal Current, or from the Pacific, or from the deep basins across the shelf break. The runoff from the continents generally moves eastward along the coast, and the continuous adding of freshwater decreases the salinity of the coastal current, and of the shelf water, from west to east. Shelf water crosses, mainly in response to local, adverse winds, the shelf break and enters the deep basins. The outflow from the shelves is generally less saline the farther downstream (east) it occurs, but in spite of this continuous export the Eurasian rivers contribute strongly to the PML also in the Canada Basin, while the runoff from North American rivers, especially the Mackenzie, flows almost directly to the straits in the Canadian Arctic Archipelago.

The Pacific Inflow

Pacific water entering through Bering Strait also contributes to the upper layers in the Amerasian Basin. The Pacific inflow crosses the Chukchi Sea shelf and reaches the Canada Basin in three streams; down the Barrow Canyon to the east, in the Central Channel east of the Chukchi Cap, and in the Herald Canyon west of the Chukchi Cap. Some water also passes through Long Strait between Wrangel Island and Siberia to the East Siberian Sea. The Bering Strait inflow has a strong seasonal cycle, being three times as strong in summer than in winter. In summer the main inflow consists of the warm ($\sim$4 °C) Bering Sea Summer Water and less saline water of the Alaskan Coastal Current that carries runoff from the Alaskan southern coast, especially from the Yukon. In winter the water on the Bering Sea shelf is cooled to freezing and the salinity increases from $\sim$32 to between 33 and 34, the most saline water coming from the Gulf of Anadyr, and the inflow consists of the Bering Sea Winter Water and sea ice.

The Pacific water is mainly confined to the Canada Basin, where it contributes to the halocline complex below the PML. The Bering Sea Summer Water is identified as a temperature maximum between 50 and 100 m, while the Bering Sea Winter Water provides a thick temperature minimum extending from 150 to 250 m (Fig. 3). This temperature minimum is also characterized by high nutrient content, indicating that the water has been in contact with the shelf bottom and entrained released nutrients from the Chukchi Sea or/and the Bering Sea shelves. Shelf water crossing the Eurasian shelf break into the Amundsen and Makarov basins may also become trapped in the Beaufort gyre and contribute to the halocline waters, primarily entering between the Pacific derived temperature maximum and temperature minimum.

The Atlantic Water

The Atlantic water enters the Arctic Ocean through Fram Strait and over the Barents Sea and In contrast to the upper layers it moves cyclonically around the Arctic basins. Its circulation was first deduced from temperature observations, which showed that the temperature of the Atlantic layer decreased cyclonically around the Arctic Ocean, being warmest north of Svalbard and coldest northeast of Greenland. Later observations have revealed boundary currents along the continental slope and separate cyclonic loops mainly following the bathymetry in the different basins. The Arctic Ocean and the Nordic Seas are characterized by closed geostrophic contours, f/H (f being the Coriolis parameter and H the depth), which allow wind driven barotropic, geostrophically balanced currents to form closed gyres, and the vorticity added by the wind field is transferred to and dissipated by the bottom torque, as the currents move around the contours.

The Barents Sea inflow branch

The Atlantic water of the Barents Sea branch separates into two major streams. The southern stream moves eastward south of the Central Bank to the large eastern depression, while the northern branch flows northward west of the Central Bank. There it splits; one stream flows east north of the bank, one stream crosses into the northern Barents Sea west of the Grand Bank, and one part returns to the Norwegian Sea as a deep, denser outflow south of Bear Island. The Atlantic water is cooled strongly in the southwestern part of the Barents Sea and as it continues between Franz Josef Land and Novaya Zemlya to the Kara Sea its maximum temperature is hardly above 0 °C. The cooling of the Atlantic water in the Barents Sea provides the largest oceanic heat input to the Arctic.

The Norwegian Coastal Current, carrying low salinity water from the Baltic Sea and runoff from the Norwegian coast, enters the Barents Sea close to the coast and becomes the Murman Current, which moves eastward until it reaches Novaya Zemlya. Here it splits with one part entering the Kara Sea through the Kara Gate between Novaya Zemlya and the Eurasian continent and the rest flowing north over the shallow banks west of Novaya Zemlya. Here the entire water column is cooled to freezing in winter and sea ice is formed. Lee polynyas are frequently present, leading to large ice production and ice export, and to the formation of brine enriched dense water on the banks. The dense water drains into the deep depression and further cools and increases the density of the Atlantic water.

The sea ice drifts over Atlantic water and starts to melt from below, creating a less saline upper layer. Sea ice, either from the banks in the Barents Sea, but also from the Kara Sea and from the Arctic Ocean, is the main source of freshwater in the northern Barents Sea and is essential for stratification and the formation of the seasonal ice cover there.

The cooled and dense Atlantic water entering the Kara Sea continues into the deep Arctic Ocean, mainly via the St. Anna Trough, where it forms colder and less saline intrusions between 1000 and 1500 m in the Nansen Basin water column, below the Fram Strait branch Atlantic water. A less dense part of the Barents Sea branch meets and mixes with Fram Strait branch water in the St. Anna Trough but remains close to the shelf break and continues eastward along the Kara Sea continental slope beside the Fram Strait branch. As the continental slope narrows north of Severnaya Zemlya, this stream moves into the basin and mixes isopycnally with the warmer Fram Strait branch, creating sharp interleaving structures (Fig. 4).

Norwegian Coastal Current water but probably also some Atlantic water remain on the Kara Sea shelf and incorporate the runoff from Ob and Yenisey. The created low salinity shelf water continues eastward and passes through the Vilkitskij Strait to the Laptev Sea, where runoff from Lena and Yana is added. Some shelf water continues to the East Siberian Sea, but the main part crosses the shelf break into the Amundsen Basin in the eastern Laptev Sea.

The Fram Strait inflow branch

The Atlantic water reaching Fram Strait in the West Spitsbergen Current enters the Arctic Ocean in two streams, one over the shelf close to Svalbard and the other moving around the Yermak Plateau, but a substantial fraction recirculates westward in the strait. The two entering streams meet east of Svalbard and move as a boundary current eastward along the Eurasian continental slope. As the Atlantic water enters the Arctic Ocean it melts sea ice and often creates an area of open water, a sensible heat polynya, in winter north of Svalbard, the Whalers' Bay. Its upper part becomes a cold, less saline layer above the warm, now subsurface core (see processes above).

The boundary current receives dense Barents Sea branch water, mainly at the St. Anna Trough but also at the Victoria Channel in the Barents Sea and at the Voronin Canyon in the Kara Sea, which intrudes into the water column below the warm Fram Strait branch core. First north of Severnaya Zemlya does the less dense part of the Barents Sea branch leave the upper slope and mix isopycnally with the warm Fram Strait branch core, reducing the Atlantic layer temperature and salinity (Fig. 4).

North of the Laptev Sea a large part of the Fram Strait branch appears to leave the slope and recirculate along the Gakkel Ridge towards Fram Strait. Colder, fresher and denser Barents Sea branch water is observed as an intermediate depth salinity minimum in the interior Amundsen Basin, indicating that also some of the Barents Sea branch leaves the slope (Fig. 4). A second splitting occurs at the Lomonosov Ridge with one part moving along ridge towards Greenland and the rest, mainly comprising Barents Sea branch water, crosses the ridge into the Amerasian Basin (Fig. 5).

In the Amerasian Basin the Atlantic water continues its cyclonic circulation. As the boundary current reaches the Mendeleev Ridge it splits, one part moves along the ridge into the Makarov Basin and the other enters the Canada Basin. At the Chukchi Cap another bifurcation takes place. One part remains at the slope and moves eastward south of the Cap, while the other part circulates around the Cap into the southern Canada Basin.

The circulation of the Atlantic water in the southern Canada Basin is debated. The generally accepted view is that the circulation remains cyclonic and that the Atlantic water moves along the slope towards the Alpha Ridge, where it again splits with one part moving into the basin towards the Chukchi Cap, forming a separate gyre in the Canada Basin. However, it has been suggested that the circulation in the southern Canada Basin is anti-cyclonic, perhaps forced by a strongly developed Beaufort Gyre. At least the upper part of the Atlantic water may then move anti-cyclonically with the overlying waters.

The boundary current at the slope crosses the Alpha Ridge and continues along the North American continental slope. There it joins the stream circulating in the Makarov Basin along the Mendeleev and Alpha ridges. At the Lomonosov Ridge the boundary current again splits with one stream crossing into the Amundsen Basin and the other following the Lomonosov Ridge. One part continues to the Eurasian continental slope, forming a closed gyre in the Makarov Basin, but some water passes through gaps in the Lomonosov Ridge to the Amundsen Basin side and flows along the ridge back towards Greenland and rejoins the boundary current.

Because the waters added to the upper layers, the Atlantic water is located too deep to leave the Arctic Ocean before it reaches Fram Strait. The only exception is the Barents Sea branch derived lower halocline (see water masses below), which partly passes through the Nares Strait and enters Baffin Bay, where it forms the characteristic Baffin Bay Bottom water with temperature −0.5 °C and salinity 34.45. The different Atlantic streams from the Amerasian and Eurasian basins meet north of Fram Strait and exit the Arctic Ocean in the East Greenland Current, which in Fram Strait is augmented with recirculating Atlantic water from the West Spitsbergen Current.

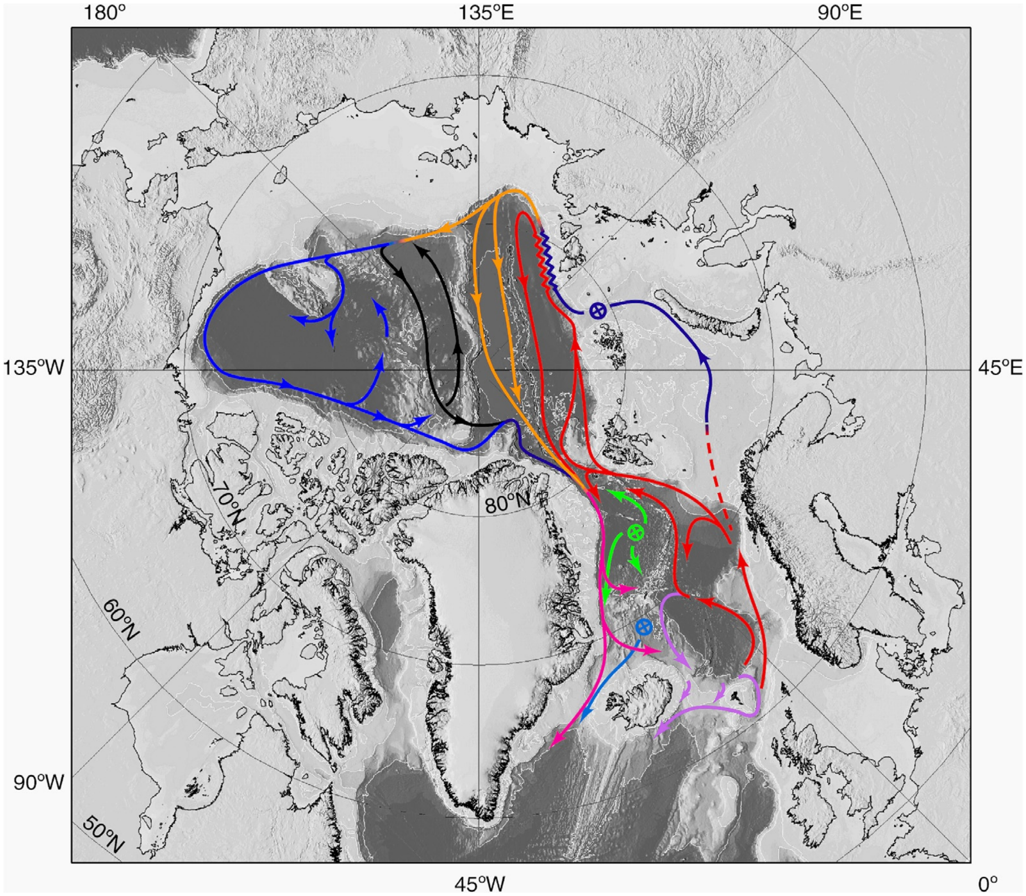

Fig. 5 Schematics showing the circulation in the subsurface Atlantic and intermediate layers of the Arctic Ocean and the Nordic Seas. The interactions between the Barents Sea and Fram Strait inflow branches north of the Kara Sea are indicated, and the colors of the different loops show the gradual cooling of the Atlantic layer. The recirculation in Fram Strait and the intermediate water formation in the Greenland Sea are shown as well as the overflows across the Greenland-Scotland Ridge. From Rudels, B., Anderson, L. G. and Eriksson, P. et al. (2012). Observations in the ocean. In: Lemke, P. & Jacobi, H.-W. (eds.) *Arctic climate change—The ACSYS decade and beyond*, pp. 117–198. Heidelberg: Springer.

Streams or Eddies?

The circulation described above might give the impression of distinct streams carrying waters around the Arctic Ocean. In reality also smaller, distinct eddies have been observed, which retain their characteristics as they are advected with the mean flow. The first eddies observed in the 1970s were located in the halocline in the Canada Basin, mostly between 100 and 170 m depth. The eddies were small, 10–20 km in diameter, predominantly anti-cyclonic with a swirling speed of 0.3 ms^{-1}, and exhibited characteristics different from those of the surrounding waters. Eddies have later been found also in the deeper layers of the water column and appear to travel significant distances without losing their identity by mixing with the surrounding water masses (Fig. 6). Recently eddy formation and eddy shedding have been evoked as a possible mechanism to balance the wind driven surface input to the Beaufort Gyre, preventing the gyre to accumulate water indefinitely.

Water Masses

The waters in the Arctic Ocean can be vertically separated into three major water masses, which all exhibit large lateral differences.

1. The PML and the halocline waters.
2. The Atlantic and intermediate waters that communicate freely across the Lomonosov Ridge.
3. The deep and bottom waters in the different basins.

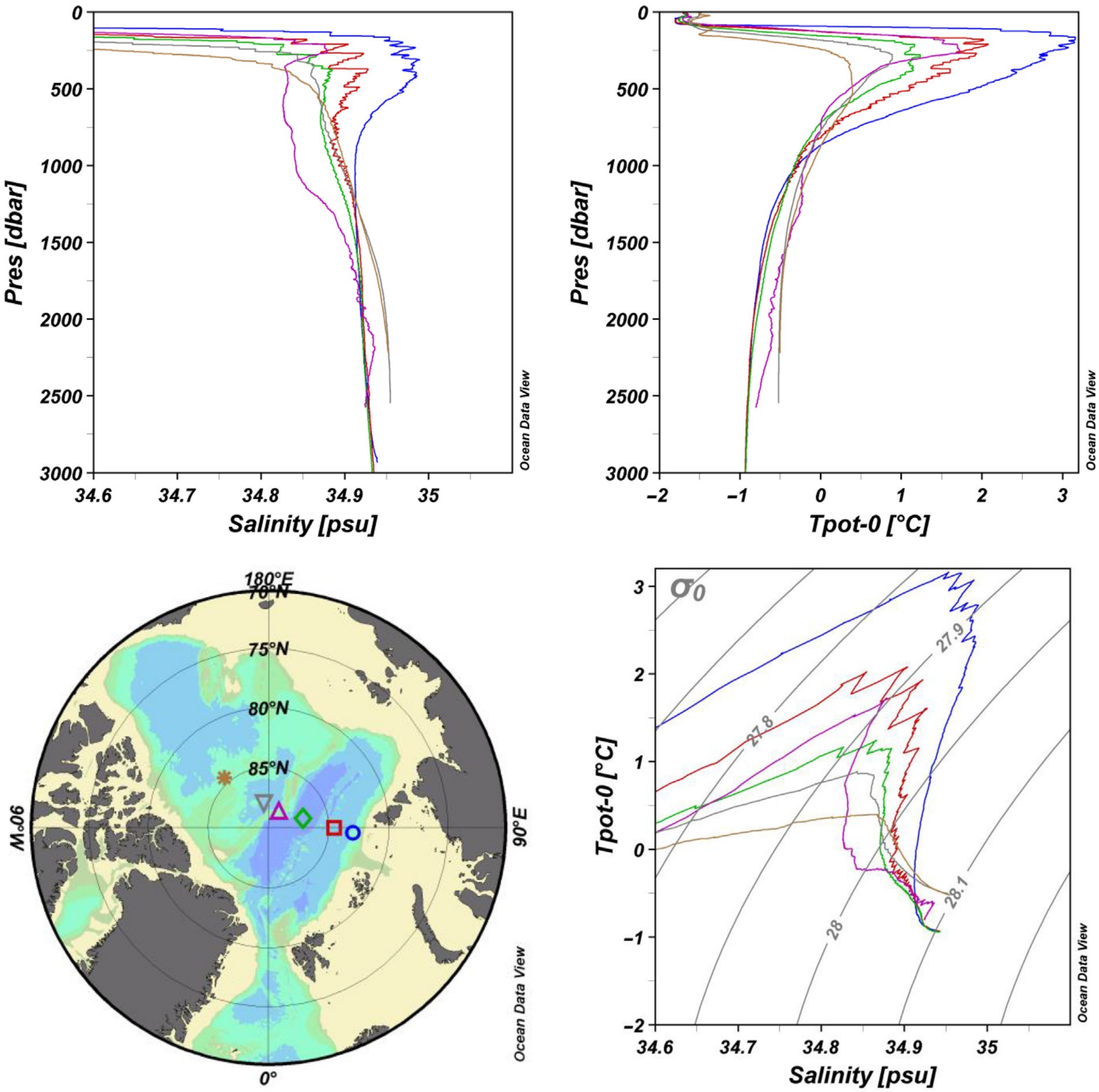

Fig. 6 Potential temperature and salinity profiles and TS curves showing the characteristics of some water masses in the Arctic Ocean: *Blue Station*; the Fram Strait branch at the Nansen Basin slope, *Red Station*; from the Gakkel Ridge showing the interleaving and cooling of the Atlantic core and the presence of less saline Barents Sea branch water at intermediate depth, *Green Station*; from the Amundsen Basin, showing a colder and less saline Atlantic layer and a stronger presence of Barents Sea branch water, *Magenta Station*; from the intra-basin in the Lomonosov Ridge showing a thick (1000 m) low salinity eddy presumably from the St. Anna inflow and a more saline layer at 2200 m depth from the Makarov Basin, *Gray Station*; from the Makarov Basin showing colder less saline Atlantic layer and warmer and more saline intermediate and deep waters, *Brown Station*; from the Alpha Ridge showing the Canada Basin water column with still colder Atlantic layer. The intermediate waters of the Amerasian Basin become warmer and more saline than the intermediate waters of the Eurasian basin between 1000 and 1200 m depth, shallower than the sill depth of the Lomonosov Ridge. Data from Polarstern 2007.

The PML actually refers to the conditions in winter, when the upper layers are homogenized by haline convection. The PML is shallowest in the Canada Basin, ~40 m, and slightly deeper in the Makarov and Amundsen basins, ~60 m. This is distinctly different from the Nansen Basin, which has the deepest, 80–100 m, and most saline, ~34.3, PML (Fig. 3). It is also different from that in the other basins because the PML here is created by sea ice melting, not by advection of low salinity water from the shelves.

In all basins sea ice melting in summer creates an about 20 m thick low salinity layer and the PML becomes stratified in salinity. As the incoming heat is used to melt ice the temperature remains close to freezing. However, in recent year a temperature maximum has been observed below the melt water layer, especially in Canada Basin. With a less compact ice cover more short wave radiation penetrates into the water column and heats the water below the melt water layer (Fig. 3). The haline stratification prevents this heat from being used for melting, and first when the melt water is removed in fall by freezing will the haline convection and mixing extend to and remove this "near surface temperature maximum".

Cold halocline waters are found below the PML, except in the Nansen Basin, where the halocline and thermocline coincide. The halocline complex is created by advection, either from the shelves or from neighboring basins (Fig. 3). The PML of the Nansen Basin becomes halocline water in the Amundsen and Makarov basins, and together with upper layer water from the Barents and Kara seas, it forms the lower halocline water in the Canada Basin. The PML from the Amundsen and Makarov basins, and the Bering Strait Summer Water and Bering Strait Winter Water all contribute to the thick upper halocline in the Canada Basin.

The Atlantic layer, defined as water with temperature above 0 °C, is in the Nansen Basin located between 150 and 750 m, but because the low salinity waters added to the upper layers the Atlantic core is displaced downward, and in the Canada Basin the Atlantic layer extends between 300 and 900 m. In the Nansen Basin the Atlantic water temperature maximum ranges between 2 and 3 °C and also displays a salinity maximum, which is not present in the other basins. The temperature and salinity are reduced by mixing between the two inflow branches and by slope convection, and the maximum temperature in the Canada Basin is between 0.5 and 1 °C (Fig. 6).

The intermediate water added by the St. Anna inflow shows up as a salinity minimum around 1000 m in the Amundsen Basin, but this minimum is not present in the Makarov and Canada basins. This is due to the input of dense water plumes from the shelves, which lowers the temperature and salinity of the Atlantic layer and slightly increases the temperature and salinity of the intermediate waters (Fig. 6).

The intermediate and deep waters beneath the Atlantic layer were earlier thought supplied through Fram Strait from the open ocean convection area in the Greenland Sea. One fact supporting this view is that the deep water in Amerasian Basin is warmer than the deep water in the Eurasian Basin. This could be explained by the presence of a ridge between the two major basins blocking the deep flow between the basins. This observation was taken as evidence for the existence of such ridge before the discovery of the Lomonosov Ridge was publicly reported. However, the Amerasian deep water is also more saline than the Eurasian Basin deep water, which can only be caused by incorporation of brine enriched plumes sinking from the shelves. These plumes also redistribute heat from the Atlantic water to the layers below, explaining the higher temperature. In fact, the water column in the Amerasian Basin becomes warmer than the Eurasian Basin water column already at 1000 m and more saline at 1200 m depth, much shallower than the sill depth of the Lomonosov Ridge (Fig. 6).

The deep waters in the Canada, Amundsen and Nansen basins have similar structure; a thick intermediate layer stratified in temperature but with salinity almost constant with depth. Around 1000 m above the bottom the salinity starts to increase, and about 200 m deeper the temperature profile has a minimum and the temperature then increases, until the deep, homogenous, 600 to 800 m thick bottom layer is reached (Fig. 7). The warmer and well mixed bottom layers have been taken as evidence of geothermal heating and stirring by convection from below. A slight increase in temperature observed in the bottom layers in recent years supports this idea. The deep temperature minimum in the Canada Basin can be explained by advection of colder water from the Makarov Basin through the passage between the Alpha and Mendeleev ridges. The minima in the Eurasian basins lie too deep to derive from an inflow through Fram Strait, and no obvious explanation for the temperature minima in the Amundsen and Nansen basins exists. One possibility could be intermittent input of dense, cold water via the St. Anna Trough, which would enter below, and thus avoid entraining, warm Fram Strait branch Atlantic water.

The Makarov Basin is different. There is no deep temperature minimum and the temperature stratification extends to the bottom layer, while the salinity becomes constant at a shallower level (Fig. 7). It has been proposed that this cold bottom layer is due to colder, fresher and denser water passing from the Amundsen Basin through deep gaps in the Lomonosov Ridge and sinking down, cooling the deep and bottom layers of the Makarov Basin. However, when the deepest known trench in the Lomonosov Ridge was explored in 2005, the densest water observed at the sill was from the Makarov Basin. This warm, saline layer extended into the Amundsen Basin, where it moved along the ridge towards Greenland and then spread out as a salinity maximum around 1500 m depth in the Amundsen Basin.

Exchanges and Transports

The exchanges between the Arctic Ocean and the world ocean have been measured by direct current observations in all passages and not only volume but also heat and freshwater fluxes have been estimated. The annual mean inflow through Bering Strait is about 1 Sv (1 Sv $= 1 \times 10^6$ m^3 s^{-1}) with large seasonal variations, less than 0.5 Sv in winter and 1.2 Sv or more in summer. The volume transport as well as the heat and freshwater fluxes appear to have increased slightly, by 10%–20%, in recent years.

The outflows of low salinity water through the three main straits in the Canadian Arctic Archipelago to Baffin Bay have from the most recent observations been estimated to 0.5 Sv (Lancaster Sound), 0.3 Sv (Jones Sound) and 0.7 Sv (Nares Strait), and the combined outflow through Davis Strait has been measured separately to 1.6 Sv. There is a decreasing trend in the reported transports and in the 2000s the combined volume transport was considered slightly above 2 Sv. The liquid freshwater export is estimated to 0.10 Sv.

Fram Strait, the deepest connection, has been monitored by a current meter array across the entire strait since 1997, but large uncertainties still prevail. The exchanges are in both direction and the estimates of the total in and outflows are about 10 Sv with a net southward flow ranging from 1 Sv to more than 2 Sv. There is a seasonal cycle. The transports are strongest in winter and weaker in summer. The presence of strong barotropic and baroclinic eddies and the large recirculation complicate the estimates of how

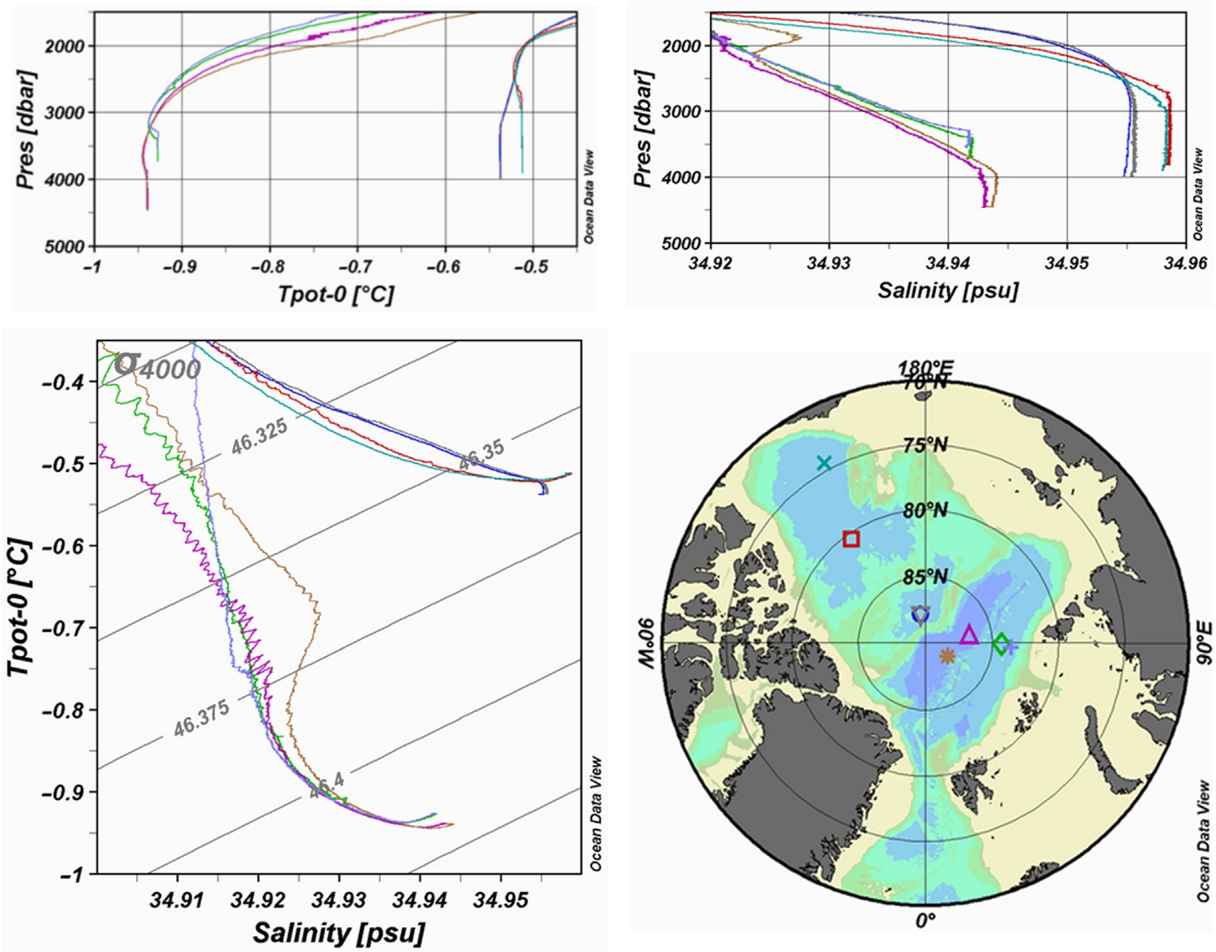

Fig. 7 Deep and bottom water characteristics from the Nansen Basin, Amundsen Basin, Makarov Basin and Canada Basin. Note the absence of a deep temperature minimum in the Makarov Basin and that the temperature minimum observed in the Canada Basin could derive from the Makarov Basin. The deep salinity maximum in the Amundsen Basin is likely be caused by Makarov Basin deep water crossing the Lomonosov Ridge. The temperature minima in the Nansen and Amundsen basins have no obvious source but could be caused by intermittent inflow of colder water via the St. Anna Trough. From Rudels, B. (2012). Arctic Ocean circulation and variability—Advection and external forcing encounter constraints and local processes. *Ocean Science* **8**, 261–286. https://doi.org/10.5194/os-8-261-2012.

much Atlantic water enters the Arctic Ocean. The eastern core, moving over the Svalbard slope carries about 1.4 Sv Atlantic water with temperature above 2 °C into the Arctic Ocean while the part deriving from the offshore branch moving around the Yermak Plateau may range from 1 to 4 Sv. The inflow of warm Atlantic water and the outflow of cold, low salinity surface water and sea ice make Fram Strait the largest heat source to the Arctic Ocean. The freshwater exports as sea ice and as liquid freshwater are about equal, 0.07–0.09 Sv. The outflow in the East Greenland Current also comprises cooled Arctic Atlantic water converging from the different loops in the Arctic Ocean, intermediate water in the 0 to −0.5 °C temperature range deriving primarily from the Barents Sea inflow branch, as well as deep waters from the Amerasian and Eurasian basins. A corresponding but weaker northward flow of intermediate and deep waters from the Nordic Seas is also present.

The second largest inflow to the Arctic Ocean occurs at the Barents Sea opening and consists of Atlantic and Norwegian Coastal Current waters. There is also a small return flow of both dense bottom water formed in the Barents Sea and less dense low salinity water. The estimates of the net inflow range from 2 to 3.5 Sv, with Atlantic water supplying 1.5–2 Sv and the Norwegian Coastal Current 0.8–1.8 Sv.

To formulate heat and freshwater budgets for the Arctic Ocean all passages must be monitored. If not, the estimates will depend upon the chosen reference temperature and reference salinity. Also when observations from all passages exist and the budgets can be determined, the transports through the different openings will still depend upon the chosen reference values.

The volume transports must balance over about a month. Longer periods of imbalance lead to a sea level slope, forcing a northward or southward barotropic flow, primarily through Fram Strait, which then removes the sea level difference. By contrast, the freshwater budget can remain unbalanced over a much longer period. The freshwater accumulation will be compensated by a change in sea level without changing the barotropic pressure below the upper layers.

The freshwater input to the Arctic Ocean is 0.124 Sv as runoff, 0.063 Sv as net precipitation over the Arctic Ocean and 0.076 Sv as liquid freshwater (relative to 34.8) from the Pacific inflow, which is balanced by an outflow of 0.10 Sv of liquid freshwater (relative to 34.8) through the Canadian Arctic Archipelago and 0.08 Sv liquid freshwater (relative to 34.8) and 0.08 Sv sea ice through Fram Strait. These values are taken as representative for the late 20th century and show the freshwater budget almost in balance. Presently the Beaufort gyre appears to accumulate freshwater, either from sea ice melt or from other freshwater sources or from a reduced outflow, indicating an imbalance in the freshwater budget.

The oceanic transports contribute to the heat lost by the Arctic to space, and the heat budget is not balanced. Several heat transport estimates have been made at single passages, especially Fram Strait, where the net southward flow makes the heat transport estimates dependent upon the chosen reference temperature. However, such efforts are not completely futile, and less formal approaches can be used to deduce the fate of the sensible oceanic heat transported through the different passages.

For example, the Bering Sea Winter Water is already at the freezing point and carries no heat and the temperature maximum of the Bering Sea Summer Water, about 0.8 Sv, is eventually reduced from 4 to $-1.5\ ^\circ\mathrm{C}$, losing practically all its heat in the Arctic Ocean. The Bering Strait heat transport then becomes about 16 TW (16×10^{12} W). The Barents Sea inflow with a mean temperature of about 5 °C is transformed into three water masses. The upper layer created by ice melting on Atlantic water, and the shelf water deriving from the Norwegian Coastal Current, perhaps 2 Sv, are both cooled to freezing temperature. The rest of Atlantic water, 1.3 Sv, supplies two water masses, the dense, cold bottom water and a warmer, less dense Atlantic core. The average temperature of these outflows would be around 0 °C, when they reach the deep Arctic basins. This leads to an heat transport of 82 TW. The upper part of the Fram Strait inflow in the West Spitsbergen Current, perhaps 1 Sv, is cooled from about 3 °C to freezing temperature as the PML in the Nansen Basin is formed, losing 20 TW. This is low compared with existing estimates and likely some additional heat, 5–10 TW, is transferred from the underlying Atlantic core to the PML in the Nansen Basin, but the main reduction in Atlantic water temperature is due to mixing with colder water, leading to no further heat loss.

The volume and the sensible heat stored in the different water masses transported by East Greenland Current can then been accounted for, and the net heat transport through Fram Strait can be further estimated by subtracting the heat transported by the East Greenland Current from the heat transported by the West Spitsbergen Current. In fact, if the mean temperature of the East Greenland Current water is taken as reference, the East Greenland Current transport no heat and the estimated northward heat transport through Fram Strait becomes a maximum. The second main oceanic heat transport, the sea ice export, ~0.08 Sv, occurs almost exclusively through Fram Strait and is around 25 TW.

Forcing

The exchanges through the separate openings have different forcing. The unidirectional barotropic inflow through Bering Strait is ultimately driven by the higher sea level in the North Pacific compared to the North Atlantic, but local sea level differences between the Bering Sea shelf and the Arctic Ocean shelves, perhaps created by different wind pattern, also contribute. In fact, strong northerly winds in winter may occasionally cause low salinity shelf water and sea ice to flow southward through Bering Strait into the Bering Sea.

The outflows through the straits in the Canadian Arctic Archipelago are mainly baroclinic and take place on the right hand side of the straits looking downstream. However, also here changes in sea level between the Arctic Ocean and Baffin Bay contribute. The Ekman transport created by westerly winds over the Arctic Ocean increases the sea level difference and the transports, while easterly wind reduces the sea level slope and the transport.

Fram Strait is more complicated. Both the West Spitsbergen and the East Greenland Current have strong barotropic components, and transports obtained by geostrophic computations are generally smaller than those found from direct current measurements, but not significantly so, and the estimates of the net outflow from direct current observations and from geostrophic computations are about equal. This suggests that the major exchanges through Fram Strait are baroclinic and that the northward and southward barotropic transports are almost in balance.

In the Barents Sea Opening the Norwegian Coastal Current enters as a buoyant boundary current, while the eastward flow of Atlantic water is predominantly barotropic and likely forced by westerly winds, creating a sea level slope between the Norwegian coast and Bear Island. There is a strong seasonality of the Atlantic water transport and in spring the net inflow occasionally vanishes.

Variability

As the first scattered observations of the Arctic Ocean water column were collected, it was assumed, perhaps by necessity, that the Arctic Ocean was a quiet ocean, at least in its deeper parts, and that observations taken in different areas and separated by decades could be combined to describe the general characteristics of the water masses and the circulation. The increase in available observations and in resolution, vertically, horizontally and in time, revealed details and variability in the Arctic Ocean just as intricate as in any of the world oceans.

The first indications of substantial changes in the Arctic Ocean were the observations of increased Atlantic layer temperatures in the Nansen Basin in 1990. This pulse of warm Atlantic water from Fram Strait was tracked as it spread through the different basins and gyres of the Arctic Ocean, largely following, and confirming, suggested circulation patterns. The change was not limited to the Atlantic layer but was also found in the PML and in the halocline waters. The shelf water, usually entering the deep Arctic basins across the Laptev Sea shelf break, continued eastward and reached the East Siberian Sea before it crossed the shelf break. The area with no halocline water expanded into the Amundsen Basin and the PML reached deeper than 120 m in winter (1996), while the absence of a halocline caused to Atlantic water to shoal. The upper layers in the Amundsen Basin and part of the Makarov Basin became more saline, while the upper layer in the Canada Basin became less saline due to accumulation of freshwater. This change was considered related to the North Atlantic Oscillation, which in the 1990s was in a strong positive state, increasing the cyclonic circulation in the Eurasian Basin.

The 1990s event was perhaps exceptional in its strength, but it was in agreement with the main variability mode in the Arctic, the Arctic Oscillation, related to the North Atlantic Oscillation. When the Arctic Oscillation index is positive, the atmospheric Polar vortex is strong and confined. The Beaufort Gyre is restricted and displaced towards southeast. When the Arctic Oscillation index is negative the Polar vortex is weaker and develops large meanders and reaches farther south. The Beaufort Gyre then expands and extends over a large part of the Canada Basin.

After the high North Atlantic Oscillation index period in the 1990s, the main shelf water outflow returned to the Laptev Sea shelf break, and the stratification and the halocline were reformed in the Amundsen Basin. Since the 1990s several pulses of warm Atlantic water have been observed passing through Fram Strait and move eastward in the boundary current in the Nansen Basin. These pulses have been traced upstream to warmer water entering the Nordic Seas from subpolar gyre. However, these later pulses appear to mainly remain in the Eurasian Basin and not cross the Lomonosov Ridge into the Amerasian Basin as did the 1990s pulse.

Change

In the recent decade the major, and most noted, change occurring in the Arctic Ocean is the reduction of the sea ice cover. Since sea ice, through its high albedo and through its isolating effect on the ocean sensible heat loss, is the one Arctic feature arguably having the largest impact on the world climate, it has become a challenge of utmost importance to determine the causes for and the future evolution of this reduction.

The most obvious cause is the observed general global warming. The Arctic radiates energy back to space, and since the global area of net heat loss is smaller than the global area with net heat gain, the temperature must rise more at high than at lower latitudes to re-radiate sufficient heat to space. The higher temperature reduces ice formation in winter and in summer a thinner and less compact ice cover implies a lower albedo and less reflected solar radiation, which increases the temperature still more.

In a warmer climate the atmospheric freshwater content and northward transport of freshwater will increase and in the Arctic more long-wave radiation will be trapped in the atmosphere, raising its temperature and the back radiation to the sea surface. A less compact ice cover also causes stronger evaporation, which further contributes to the atmospheric moisture content.

However, the observed temperature increase and recently reported shoaling of the Atlantic layer, especially in the Nansen Basin, have also been suggested as possible causes for the diminished ice cover. The Atlantic layer in the Nansen Basin is only separated from the ice cover by the PML, and if the Atlantic temperature is higher, more heat can go to ice melt. However, more melting increases the stratification, leading to less entrainment and a shallower PML, reducing the heat input. The Atlantic water comes closer to the ice cover, but a warmer Atlantic layer only adds marginally more heat to ice melt. Actually, the fact that the different warm pulses can be followed around the Eurasian Basin rather suggests that once the PML is formed, most of the heat remains in the Atlantic layer and is returned to the Nordic Seas through Fram Strait.

A more drastic effect of a higher temperature in the Atlantic inflow could occur in the Barents Sea. Warmer Atlantic water reduces the ice formation over the shallow banks in the central and eastern Barents Sea, and if less ice is formed and drifts northward, the freshwater input to the northern Barents Sea is reduced, leading to weaker stratification. This affects the ice formation and the seasonal ice cover and might contribute to the observed northward retreat of the ice edge, which shows up as a large surface air temperature anomaly, going from below $-20\ ^{\circ}\mathrm{C}$ to freezing temperature. If this also leads to a substantially larger heat input from the ocean to the atmosphere in the northern Barents Sea is a more difficult question to answer.

The Pacific inflow also shows large temperature variations and years with warm and strong inflow will affect the ice formation in the Chukchi Sea because more sensible heat must be extracted from the water column before it reaches freezing temperature and sea ice can form.

The Arctic Ocean freshwater budget, inputs, exports and balances, has come into focus as the perhaps major component influencing and connecting the Arctic and the global climate. The freshwater export in buoyant boundary currents is equal to that of a pure freshwater layer. If the freshwater budget is balanced and the sources and the ice export are known, the freshwater storage in the Arctic Ocean can be estimated by assuming that the thickness of the freshwater layer in the passages is equal to the average thickness of the freshwater layer in the Arctic basins. This does not take into account the effects of the wind fields over the Arctic Ocean, which may accumulate (disperse) the upper layer water in the central parts of the basins and reduce (increase) the thickness and the transports through the gateways.

Recently the Arctic Ocean, especially the Beaufort Gyre, has accumulated freshwater due to the prevailing atmospheric circulation characterized by a negative Arctic Oscillation index, and when (if) this freshwater is released, the thickness of the upper layers in the gateways and the outflow of liquid freshwater will increase, but dramatic outflow events may not occur. This is more likely to happen with an anomalously large sea ice export, which is not constrained by geostrophy as is the liquid freshwater export. If the ice cover continues to diminish, such events are not expected.

If the ice export is reduced, it is probable that relatively more freshwater exits through the Canadian Arctic Archipelago than through Fram Strait. However, should glacial ice melt from Greenland increase significantly, the salinity of the upper layer in Baffin Bay would decrease, which may reduce the baroclinic transport through the straits in the Canadian Arctic Archipelago and shift it towards Fram Strait.

While the low salinity upper layer is crucial for the maintenance of the ice cover, it inhibits deep convection, and one major, global concern, is that the increased atmospheric freshwater input to the Arctic Ocean, caused by a warmer climate, could affect the Atlantic Meridional Overturning Circulation (AMOC). This question was investigated by Henry Stommel in 1961, using a simple conceptual box model with heating and evaporation at low latitudes and cooling, precipitation and sinking at high latitudes. The model shows the strong direct circulation with sinking and convection at high latitudes and a returning deep flow, but also that larger freshwater input at high latitudes can lead to a reversed circulation with sinking at low latitudes and a southward flow of low salinity water at the surface. This reversed circulation is weaker, because heat is not directly lost to the atmosphere and mechanical energy is required to mix water from below into the surface layer.

Presently we have both situations; cooling and sinking in the subpolar gyre and in the Nordic Seas, and stratification and mechanical mixing in the Arctic Ocean. Less ice formation would increase the stability, and less water would be mixed into the surface layer. However, if the ice formation is reduced not enough sea ice might drift into the Nansen Basin to create and maintain the PML throughout the winter, and convection might here break through into the Atlantic layer, creating overflow water rather than Polar Surface water.

The Greenland Sea presently produces less dense water than before. No local deep water renewal takes place, and the deeper layers have gradually been replaced by Arctic Ocean deep waters, making the Greenland Sea deep waters warmer and more saline. The Arctic intermediate water now formed in the Greenland Sea has also become warmer and more saline, but it is dense enough to supply overflow water to the North Atlantic and the AMOC.

One factor that contributes to a sustained convection in the Nordic Seas and in the subpolar gyre also with larger freshwater input is that the liquid freshwater outflow is confined to the boundary current and seldom enters the convective gyres. Sea ice is different. It can be driven by the winds to the convecting gyres and perhaps shut them down temporarily. Such shutdowns might have taken place during glacial epochs, e.g. during the Dansgaard-Oeschger events. Presently, a shutdown of the convection caused by ice drift is not likely, neither in the Greenland Sea nor in the subpolar gyre.

The oceanic heat carried to the Arctic Ocean is but a small fraction (5%) of the atmospheric heat transport. Should the atmospheric heat transport increase then no ice export, and perhaps no oceanic heat transport, would be needed to supply the heat radiated to space to balance the global heat budget. There would be ice formation in winter, but the ice cover would disappear in summer and the ice export would be negligible. Should the AMOC be reduced, or even reversed, the diminishing of the northward oceanic heat transport could cause cooling especially in northwestern Europe but it would have less effect in the Arctic. Perhaps the cyclonic circulation in the Eurasian Basin would be reduced if the low pressure systems from the North Atlantic do not reach into the Arctic and the anticyclonic circulation centered over the Beaufort gyre would expand. These thoughts are, however, mere speculations, and global climate models are required to provide more detailed and realistic scenarios.

Suggested Reading

The classical reference for the Arctic Mediterranean Sea is Coachman and Aagaard (1974). Comprehensive, recent overviews, where original references can be found are Bluhm et al. (2015), Mauritzen et al. (2013), Rudels (2012, 2015), Rudels et al. (2012). The exchanges between the Arctic Ocean and the world ocean are reviewed by Beszczynska-Möller et al. (2011) and Tsubouchi et al. (2012). The first indication of major changes in the Arctic Ocean water column was reported by Quadfasel et al. (1991). The effects of the oceanic heat transport are discussed in Polyakov et al. (2010, 2012, 2017) and Rudels et al. (2013, 2015). Freshwater transports and budgets are summarized in Dickson et al. (2007), Haine et al. (2015) and Carmack et al. (2016). The numerical examples given in the text mainly use Haine et al. (2015) and Rudels et al. (2015). The dynamics of the freshwater export and the interactions between the Atlantic inflow and sea ice are examined in Rudels (2010, 2011, 2016), and the role of the Arctic Ocean in the global climate is well described in Aagaard and Carmack (1994). The references to the classical work are Nansen (1902) and Stommel (1961).

The present article is mainly descriptive and as introduction to more theoretical and modeling work Nøst and Isachsen (2003) and Spall (2012, 2013) as well as the Forum for Arctic Modeling and Observational Synthesis (FAMOS) website https://famosarctic.com could be consulted.

Acknowledgment

Ocean Data View (Schlitzer, 2017) has been used to construct some of the figures.

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