ARCTIC BASIN CIRCULATION

B. Rudels, Finnish Institute of Marine Research, Helsinki, Finland

Copyright © 2001 Academic Press

doi:10.1006/rwos.2001.0372

Introduction

The Arctic Ocean, the northernmost extension of the North Atlantic, is an almost land-locked sea. It is connected to the Nordic Seas through the 2600 m deep, 500 km wide Fram Strait and over the extensive, ~ 250 m deep, Barents Sea, and to Baffin Bay through the shallow (150–250 m), narrow straits in the Canadian Arctic Archipelago. The last two passages are usually considered parts of the Arctic Ocean. The Arctic Ocean communicates with the North Pacific through the 50 m deep and 50 km wide Bering Strait, forming a direct link between the two opposite poles of the world ocean, the low salinity North Pacific and the high salinity North Atlantic.

The wide, shallow shelves of the Barents, Kara (50–100 m), Laptev (<50 m), East Siberian (<50 m), and Chukchi (~ 50 m) Seas comprise about 1/3 of the area of $9.4 \times 10^{12}\,m^2$. The rest consists of two main basins, the Eurasian and the Canadian Basins, with maximum depths > 4200 m and > 3900 m, respectively. They are separated by the Lomonosov Ridge (typical sill depth 1700 m) that runs from Siberia to Greenland, passing close to the north pole. The Eurasian Basin is further divided into the Amundsen and Nansen basins by the Nansen-Gakkel Ridge, and the Canadian Basin is separated into the Canada and Makarov basins by the Alpha-Mendeleyev Ridge (**Figure 1**).

The severe, high-latitude climate of the Arctic determines the oceanography of the Arctic Ocean. There is a net heat loss to space, and heat is supplied from lower latitudes, primarily by the atmospheric circulation but also by the northward advection of warmer water, mainly entering the Arctic Ocean through Fram Strait. The atmosphere also transports fresh water from lower to higher latitudes. The extensive catchment area of the Arctic Ocean leads to a large (0.1–0.15 Sverdrup (Sv); $(1\,Sv = 10^6\,m^3\,s^{-1})$), seasonally varying runoff that creates a strong stability in the upper part of the water column and limits the winter convection to 50–100 m. This upper layer is cooled to freezing temperature, and ice is formed. In winter, sea ice covers the entire Arctic Ocean except the Atlantic-dominated south-western Barents Sea. In summer, the ice in the marginal seas disappears, but an ~ 3 m thick ice cover remains throughout the year in the central Arctic Ocean. The ice export (~ 0.1 Sv), mainly takes place through Fram Strait ($> 90\%$), and constitutes a transport of latent heat which, arguably, is the largest advective contribution to the Arctic Ocean heat balance.

Owing to difficult observation conditions the Arctic Ocean is one of the least explored oceans. After the 1893–1896 Fram expedition, work in the high Arctic was for a long time limited to a few drifting stations, and to airborne expeditions, mostly carried out by the former Soviet Union. In the 1980s and 1990s ice breaker and submarine expeditions have added greatly to the existing database.

The Inflow of Atlantic and Pacific Waters

The waters of the Arctic Ocean are supplied mainly from the North Atlantic. The inflow consists of two branches, entering through Fram Strait and over the Barents Sea (**Figure 2**). A large fraction of the Atlantic Water (AW) in the West Spitsbergen Current, 1–2 Sv, recirculates in Fram Strait, but 1–3 Sv enters the Arctic Ocean in two streams, one north of Svalbard and another, weaker, west and north of the Yermak Plateau. North of Svalbard the warm (3–5°C) AW encounters, and melts, sea ice, and its upper part is transformed into a less saline layer that flows above the AW core eastward along the continental slope. The approximate numbers given for the transports reflect not only the uncertainty of the estimates but also the recently recognized large variability of the circulation.

The branch entering the Barents Sea comprises both AW and less saline water of the Norwegian Coastal Current. The salinity of the AW is reduced by net precipitation, and by mixing with the coastal waters to the south and with less saline surface water and sea ice to the north. About 1/3 recirculates in the western Barents Sea and returns to the Norwegian Sea, mostly as colder and denser water. The rest, 2–3 Sv, continues northward and eastward, and the main part passes into the Kara Sea, primarily between Franz Josef Land and Novaya Zemlya. The two inflow branches have a common

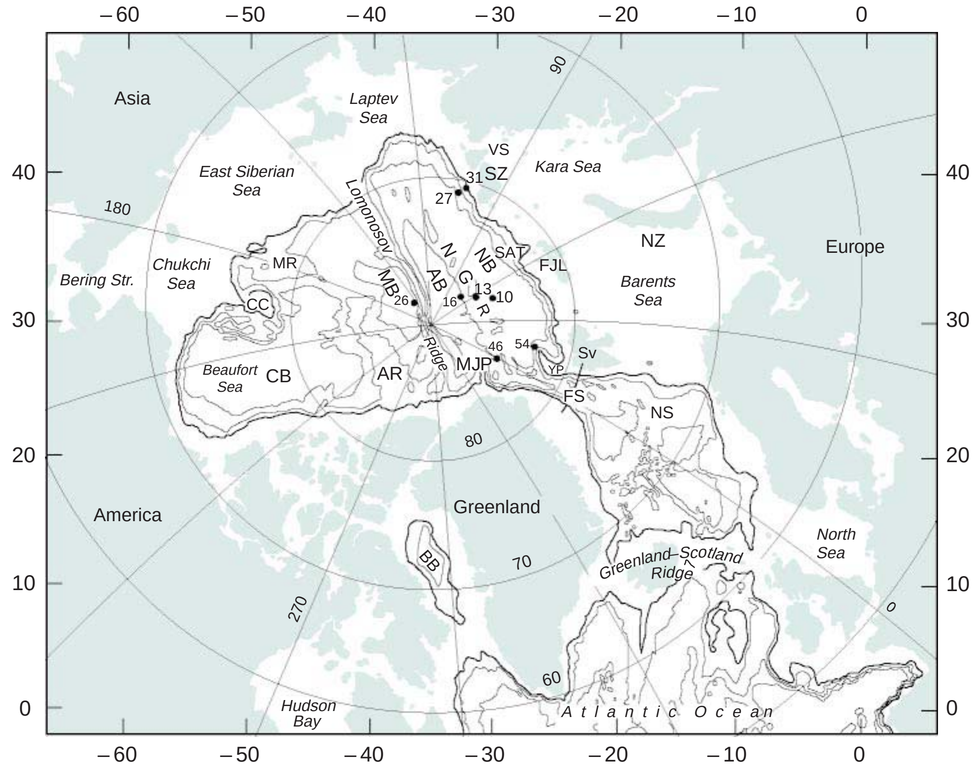

Figure 1 Map of the Arctic Mediterranean Sea showing geographical and topographical features. The positions of the stations presented in **Figures 4** and **5** are given as black dots and the section in **Figure 7** as a black line. Two coordinate systems are used. The map shows the latitudes and longitudes of the geographical system. The map is a Mercator projection with the positive pole at 30°N; 115°E (China). The zero meridian passes through Greenwich, and the 'longitude' is counted negative to the left. Distances relate in the customary way to the map coordinates (latitudes), given in degrees on the frame. The North Atlantic with the Arctic Mediterranean Sea shows small distortions in this system in comparison to more common presentations. Bottom contours are shown by isolines for 1000, 2000, 3000, and 4000 m of decreasing thickness, AB, Amundsen Basin; CB, Canada Basin; MB, Makarov Basin; NB, Nansen Basin; AR, Alpha Ridge; MR, Mendeleyev Ridge; NGR, Nansen-Gakkel Ridge; CC, Chukchi Cap; MJP, Morris Jesup Plateau; YP, Yermak Plateau; BB, Baffin Bay; FS, Fram Strait; NS, Nordic Seas; SAT, St Anna Trough; VS, Vilkiltskij Strait; FJL, Franz Josef Land; NZ, Novaya Zemlya; Sv, Svalbard; SZ, Severnaya Zemlya.

source, AW crossing the Greenland-Scotland Ridge, and their transports are likely to vary out of phase.

The inflow of Pacific Water (PW) through Bering Strait is about 0.8 Sv with a strong seasonal cycle, 1.2 Sv in summer and 0.4 Sv in winter. It is driven by the higher sea level in the Pacific than in the Arctic Ocean. The PW is less saline than the AW, 32.5 PSU as compared to 35 PSU, and less dense. It provides water to the Chukchi Sea and to the upper layers in the Canadian Basin.

Circulation

In the ice-covered part of the Arctic Ocean the circulation of the surface water has been deduced from sea ice drift. It is primarily wind-driven. The high-pressure cell above the Beaufort Sea creates an anticyclonic gyre that feeds the Transpolar Drift, crossing the Lomonosov Ridge from the Canadian to the Eurasian Basin close to the pole. A second ice stream originates from the Siberian shelves, especially the Laptev Sea, and joins the transpolar drift north of Fram Strait. The Siberian Branch was the stream carrying Fram (**Figure 2**). These motions agree with the dynamic topography, which shows the accumulation of less dense surface water in the Beaufort Sea. Eastward flowing coastal currents are present on the shelves.

The anticyclonic flow is restricted to the uppermost layer, and the deeper circulation is dominated by a subsurface boundary current that flows along continental slope cyclonally around the basins. It splits at prominent topographic features, and streams follow the ridges into the interior, while the motions over the abyssal plains are weak (**Figure 3**). The boundary current starts as the Fram Strait branch, including its less saline upper part, flows eastward along the Eurasian continental slope. It is augmented by outflows from the Barents Sea, entering at and below its warm core. A part of the

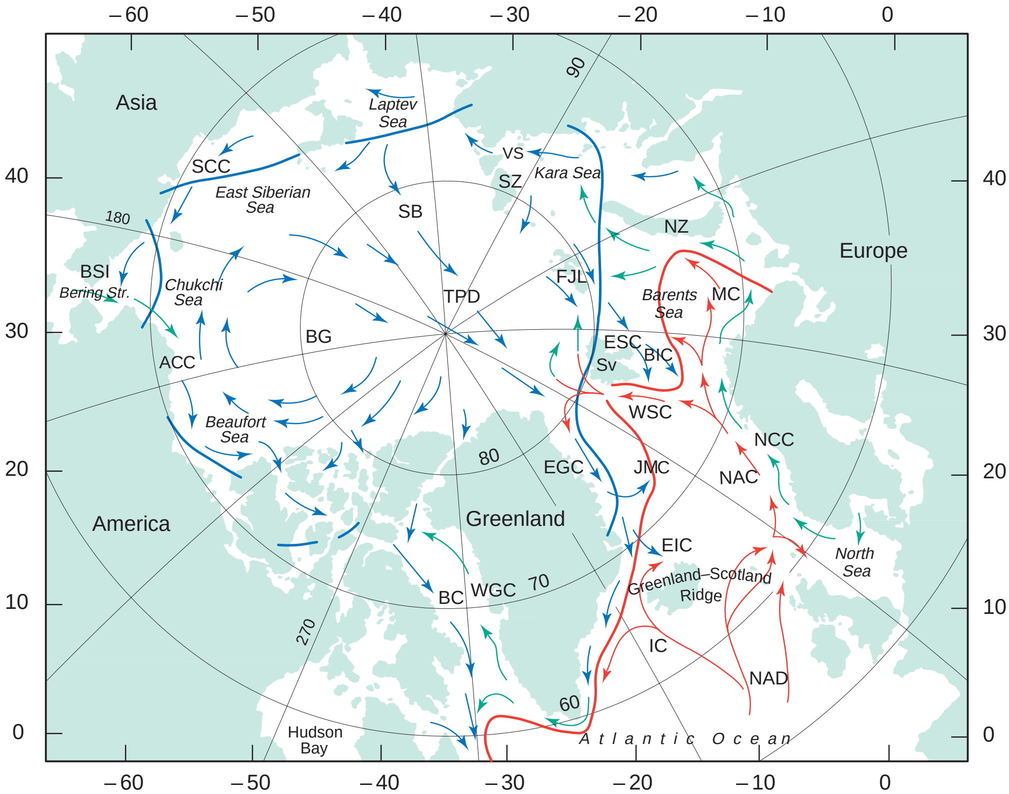

Figure 2 The circulation of the surface waters. Warm Atlantic currents are indicated by red, cold Arctic currents by blue, and less-saline and/or transformed currents green arrows. The average maximum extent (red line) and the average minimum extent (blue line) of the ice cover are also shown. ACC, Alaskan Coastal Current; BC, Baffin Current; BIC, Bear Island Current; BG, Beaufort Gyre; BSI, Bering Strait Inflow; EGC, 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; SCC, Siberian Coastal Current; TPD, Transpolar Drift; WGC, West Greenland Current; WSC, West Spitsbergen Current.

boundary current enters the deeper troughs in the northern Barents Sea, and the St Anna Trough in the Kara Sea. Barents Sea branch water reaches the Arctic Ocean and joins the boundary current as a strong subsurface inflow at the eastern part of the St. Anna Trough, and the two branches continue eastward; the Barents Sea branch closest to the slope.

The mixing at the front between the two branches creates interleaving layers, and the profiles show inversions in temperature and salinity, indicating the possibility of double-diffusive convection (**Figure 4**). Similar, more extensive layers have been observed in the interior of the basins (**Figure 5**). It is not known whether double-diffusive convection could drive the waters in the layers, allowing them to expand, and thus contribute to the lateral mixing in the interior of the basins, or whether the layers are formed, and all but run down, in active frontal zones. Their wide distribution would then be due to the advection, with the mean circulation of fossil, almost inactive, layers.

A fraction, consisting predominantly of Fram Strait branch water, becomes separated from the boundary current and returns toward Fram Strait within the Nansen Basin. North of the Laptev Sea lateral mixing has all but removed the front between the two branches, and Fram Strait branch water is again present at the continental slope. At the Lomonosov Ridge half the current, perhaps more, turns northward, bringing Fram Strait as well as Barents Sea branch water along the ridge into the interior Amundsen Basin. The rest enters the circulation (**Figure 3**). Barents Sea branch water has been identified by colder, less saline anomalies in the Makarov Basin deeper than 1000 m, indicating that the boundary current extends down to sill depth. The boundary current again splits at the Mendeleyev Ridge, and a further split may occur at the Chukchi Cap. The boundary current continues along the American continental slope, splitting at the Alpha Ridge and at the Lomonosov Ridge, possibly forming gyres in the Canada and the Makarov Basins. The boundary current returns to the

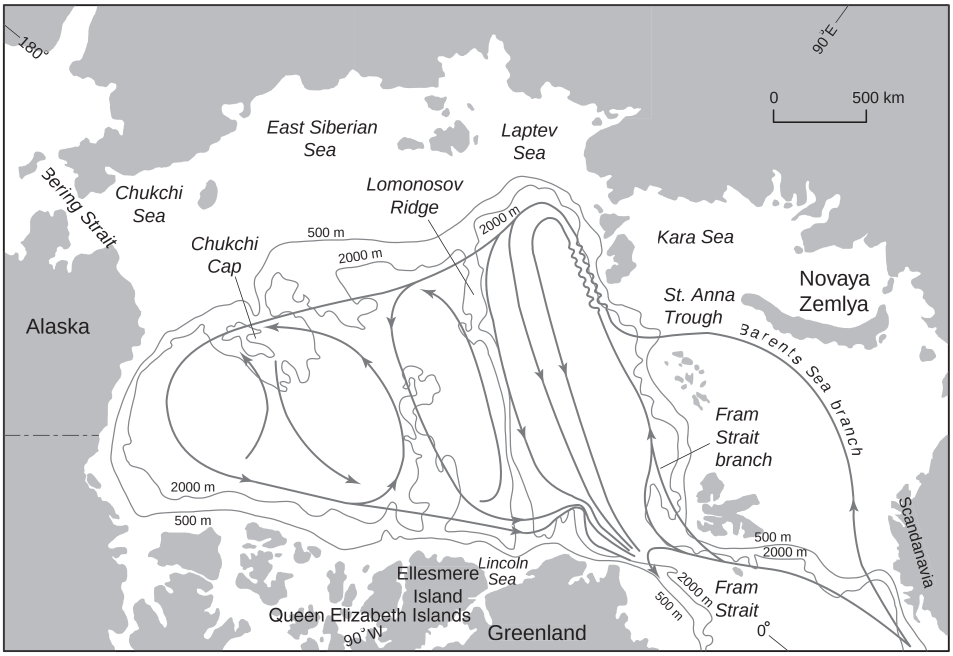

Figure 3 Schematic indicating the circulation of the Atlantic and intermediate waters and the mixing of the two inflow branches north of the Kara Sea. (Adapted from Jones EP, Rudels B and Anderson LG (1995). Deep waters of the Arctic Ocean: Origins and Circulation. *Deep Sea Research I* 42: 737–760.)

Eurasian Basin north of Greenland and continues toward Fram Strait, merging with the streams exiting the Eurasian Basin into the East Greenland Current.

The inflows through Fram Strait and the Barents Sea suggest a boundary current transport of $\sim 4\,\mathrm{Sv}$. Perhaps 3 Sv recirculate within the Eurasian Basin, while 1 Sv enters the Canadian Basin. Few current measurements exist, bur recent observations west of, at, and east of the Lomonosov Ridge showed velocities of $0.05\,\mathrm{m\,s^{-1}}$ in the boundary current, and the transport was estimated to 5–6 Sv north of the Laptev Sea. About half flowed north along the Lomonosov Ridge and half entered the Canadian Basin. These larger transports imply either that the inflow estimates are too low or that the boundary current also includes a circulation internal to the basins.

The observed velocities often exceeded the mean, indicating the presence of low-frequency waves and/or eddies. Anticyclonic eddies have been reported earlier, mainly from the Canadian Basin pycnocline. They are 10–20 km in diameter highly energetic with maximum velocities above $0.3\,\mathrm{m\,s^{-1}}$, and associated with anomalous water properties. The eddies are expected to survive longer than a year, and in the Canadian Basin a possible formation process is the injection of water from Bering Strait into the pycnocline, the inflow becoming unstable, creating eddies.

Water Mass (Trans)Formation

Low-salinity water is formed on all shelves in summer, when the river runoff occurs, and is exported to the deep basins. The resulting stability of the upper layers limits the convection in the interior Arctic Ocean to 50–100 m and the properties in the deeper layers are determined by advection. On the shelves, ice formation and brine rejection lead to convection and to accumulation of saline water at the bottom. The salinity increases throughout the winter. The density reached depends upon the heat loss, the initial salinity, the depth of the shelf, and the existence of polynyas. These are mainly located close to coasts and islands like Svalbard, Novaya Zemlya, Franz Josef Land, and Severnaya Zemlya, where offshore winds maintain an open free sea surface and a large ice production.

As the dense water crosses the shelf break, it sinks down the continental slope as thin, intermittent boundary plumes. The plumes have initial temperatures close to freezing, but their salinity varies. Less-saline plumes enter and cool the AW in the boundary current. More saline, denser plumes sink through the AW, entraining and redistributing warmer AW downward. The temperature of the plumes increases, and their density anomaly decreases. When their density matches that of the ambient water column, they merge with the surroundings.

For plumes to sink, bottom friction must be strong enough to break the balance between the buoyancy and the Coriolis accelerations. This is consistent with a high turbulence level and large entrainment into the plumes. Most plumes observed in the Arctic Ocean have been fairly shallow, and colder and less saline than their surroundings. Only west of Svalbard and north of Severnaya Zemlya have plumes more saline than their surroundings been observed, located deeper than 2000 m (**Figure 4D**). The properties of the plumes in the Arctic Ocean have thus been deduced mainly from their effects on the characteristics of the intermediate and deep waters in the different basins, not by direct observations.

A separator process that transforms a water mass into less dense as well as denser waters is characteristic of the Arctic Ocean, and it strongly affects the waters of the Barents Sea branch. Over the shallow parts of the Barents Sea, especially west of Novaya Zemlya, the water column is cooled to freezing. Ice is formed and, by favorable winds, driven from the coast, keeping the sea surface open and the ice production high. Brine released during ice formation increases the density of the underlying water, which eventually sinks into the deeper depressions, cooling

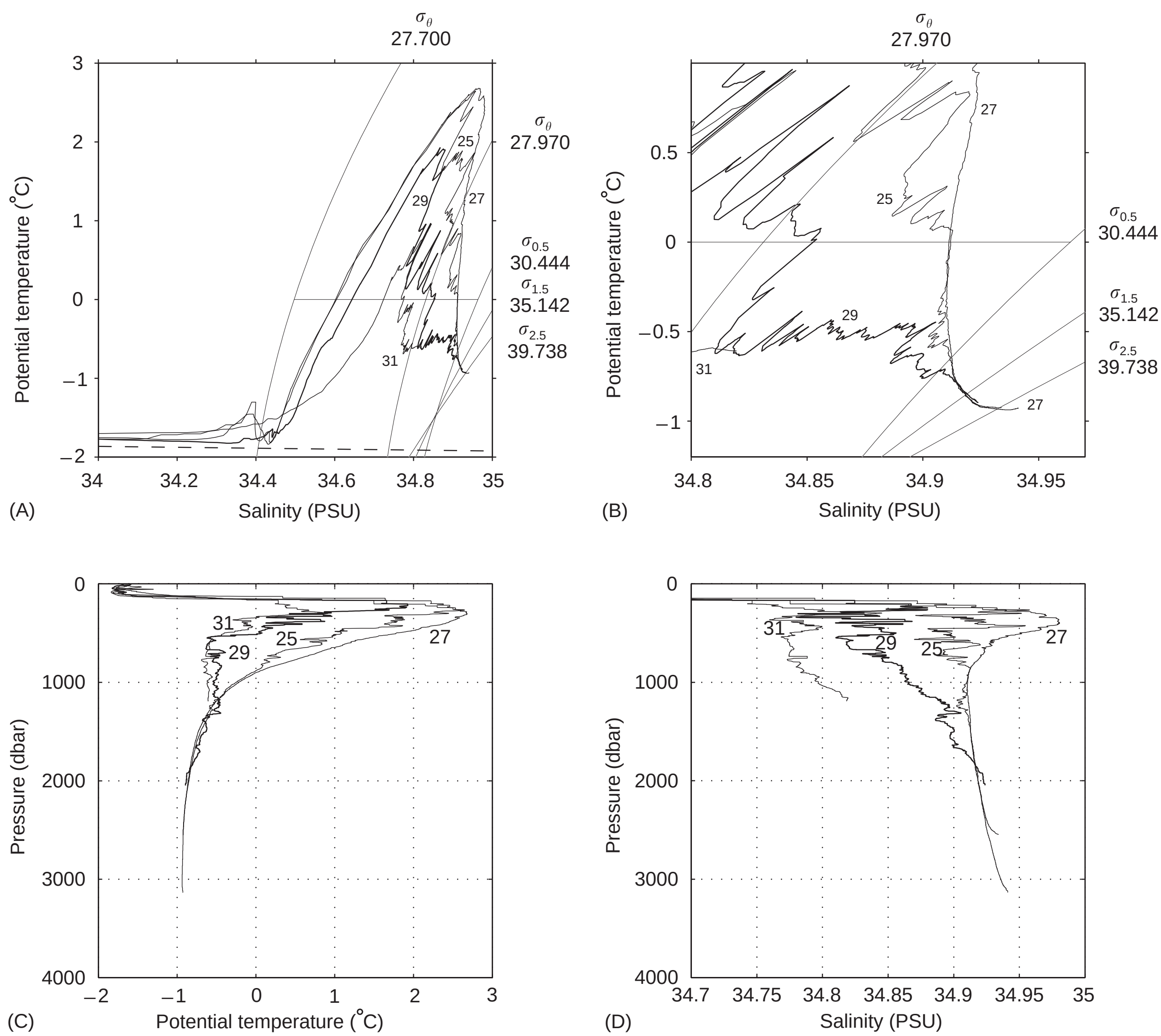

Figure 4 (A) Potential temperature salinity (– *S*), (B) Θ–*S* curves (blow-up), (C) potential temperature profiles, and (D) salinity profiles for stations 25, 27, 29, and 31 taken in 1995 by R.V. Polarstern north of Severnaya Zemlya is the potential density referred to the sea surface and $\hat{x}$ is the potential density referred to $\times 10^3$ dbar. The colder Barents Sea branch occupies the slope with the less-saline, less-dense water closest to the shelf, while the Fram Strait branch is displaced from the slope. The mixing between the two branches creates inversions in the Atlantic and intermediate layers. The small salinity increase seen at the bottom on stations 29 and 25 (D) could indicate deep, saline slope plumes.

the main inflow and increasing its density. The ice eventually melts, partly because of an upward heat flux from the AW, partly because of the warming in spring and summer, and creates a less-dense surface layer.

The waters of the Barents Sea branch, passing to the northern Barents Sea, and eastward into the Kara Sea, thus become colder and less saline, but denser as well as less dense. The denser waters continue down the St Anna Trough. The densest

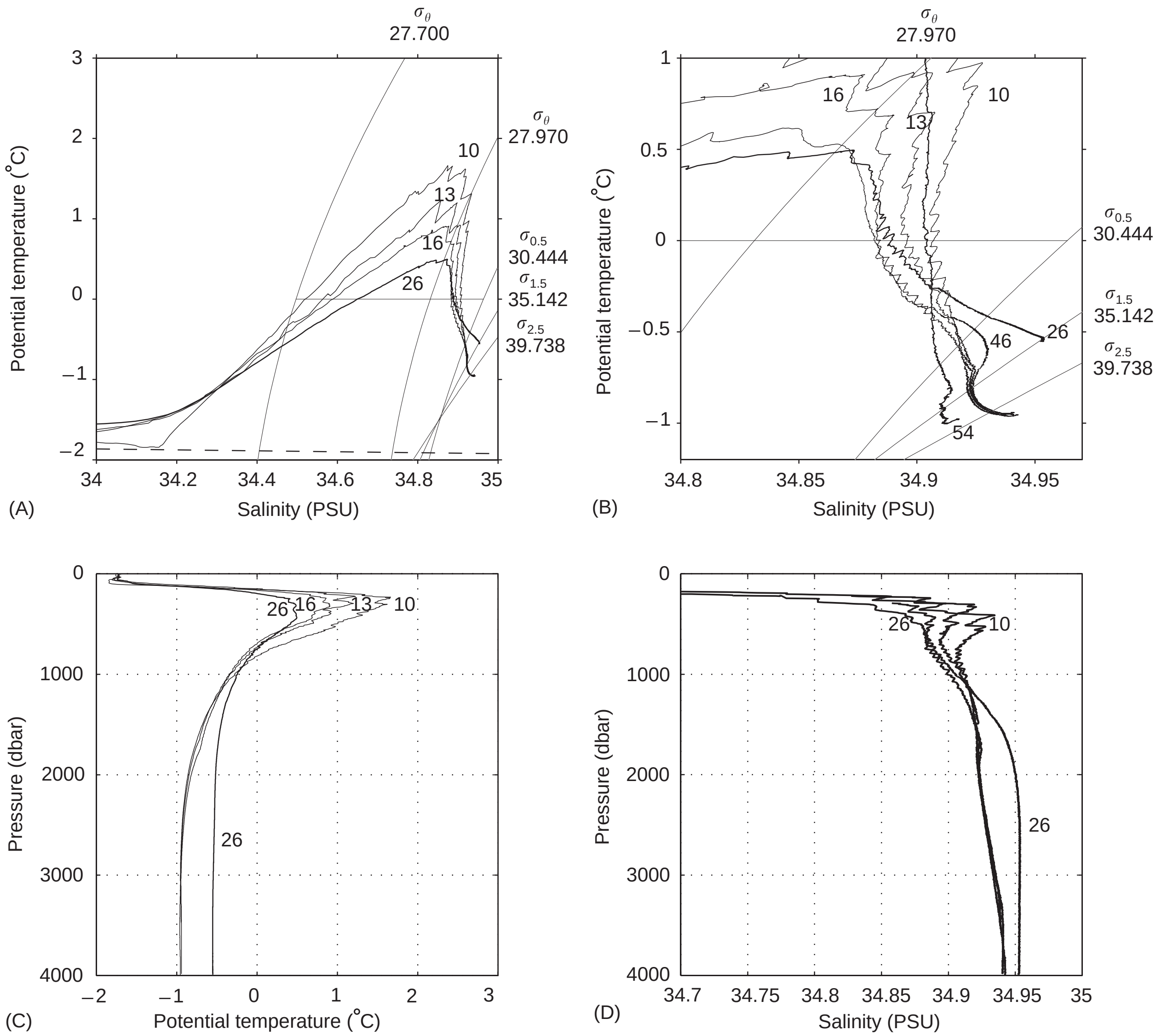

Figure 5 (A) Potential temperature salinity ($-$ *S*), (B) Θ–*S* curves (blow-up), (C) potential temperature profiles and (D) salinity profiles for stations 10, 13, 16, 26, 46 (only in (B)), and 54 (only in (B)) taken in 1991 by IB Oden is the potential density referred to the sea surface and $\hat{x}$ is the potential density referred to $\times 10^3$ dbar. The isopycnal $\sigma_\theta = 27.70$ separates the Polar Surface Water from the intermediate layers and the isopycnal $\sigma_{0.5} = 30.444$ separates the intermediate and deep waters. The $\sigma_{1.5} = 35.142$ isopycnal is roughly at the sill depth of the Lomonosov Ridge and the $\sigma_{2.5} = 39.738$ isopycnal lies at the sill in Fram Strait. The temperature and salinity of the Atlantic Layer and the uPDW decrease from the Nansen-Basin (station 10) to the Nansen-Gakkel Ridge (station 13) and the Amundsen Basin (station 16), indicating the Barents Sea branch water becomes more prominent and the Fram Strait branch water less prominent in the recirculation branches lying closer to the Lomonosov Ridge. Inversions and layers extend across the interior of the Eurasian Basin. The colder, less-saline Atlantic Layer and the warmer, more-saline uPDW in the Canadian Basin are seen on station 26. In (B) station 26 reveals the constant salinity and the slight lowering of temperature with depth in the CBDW, while stations 10, 13, and 16 show the colder EBDW with a salinity maximum at 1700 m and the salinity increase toward the bottom. The intermediate salinity maximum has the same density as the CBDW and is stronger closer to the Morris Jesup Plateau (station 46). The Arctic Intermediate Water (AIW) and Nordic Sea Deep Water (NSD) entering through Fram Strait are colder and less saline than EBDW and not dense enough to affect the deepest layers of the Eurasian Basin (station 54). They are colder, more saline and denser than the Barents Sea inflow branch (compare **Figure 4B**).

part enters deeper than 1000 m, adding colder, less saline water at intermediate depth without significant entrainment (**Figure 4**). The less saline upper layer remains on the shelf, incorporating the runoff from Ob and Yenisey. The main part then passes through the Vilkiltskij Strait into the Laptev Sea, where the runoff from the third large Siberian river, Lena, is added. Not until this Atlantic-derived shelf water reaches the eastern Laptev Sea does a significant amount cross the shelf break and contribute to the low-salinity Polar Mixed Layer (PML), which is present over most of the interior Arctic Ocean. The rest stays on the shelf and enters the East Siberian Sea.

Water Masses

No generally accepted water mass classification exists for the interior Arctic Ocean, but three major layers should be recognized: (1) the upper low-salinity Polar Surface Water (PSW) where the fresh water input dominates the transformation of the inflowing waters; the PSW consists of the PML and the halocline; (2) the intermediate waters, comprising the Atlantic Layer and the upper Polar Deep Water (uPDW), which are mainly derived from the two inflow branches and the boundary current; (3) finally, the deep waters, largely formed by slope convection in the Arctic Ocean. The layers are here separated by the $\sigma_\theta = 27.70$ and $\sigma_{0.5} = 30.444$ isopycnals, respectively. The upper isopycnal loosely marks the change from a halocline to a thermocline in the water column, and the deeper isopycnal approximately separates waters with properties predominately affected by the inflows from waters with characteristics created by slope convection (**Figures 4** and **5**).

The uppermost water mass is then the low-salinity, 30–50 m thick PML, homogenized locally in winter by freezing and haline convection. In summer, ice melt creates a 15–20 m thick top layer with still lower salinity. The PML is supplied by the outflow from the shelves in summer, but also the Bering Strait inflow contributes low-salinity water. The salinity of the PML varies throughout the Arctic Ocean, being lower in the Canadian than in the Eurasian Basin (30–31 PSU compared to 32–33 PSU).

The halocline lies between the PML and the thermocline. It is not just a transition layer but a distinct water mass with a volume three times that of the PML. Its temperature remains close to freezing, while its salinity increases rapidly with depth. This excludes its formation by direct vertical mixing between the PML and underlying warmer water and implies advective sources. In the Canadian Basin the halocline is thicker (200–250 m) than in the Eurasian Basin (100–150 m) and has several sources. In addition to brine-enriched water from the shelves, both the warmer Bering Strait Summer Water (Salinity ~ 32.4 PSU) and the cold and more saline Bering Strait Winter Water (33.1 PSU) have been identified. Another, distinctly Canadian Basin, feature is a cold nutrient maximum with salinity 33.1 PSU, commonly found at 80–90 m, which has been traced to the Chukchi shelf. The high nutrient content indicates that winter convection in the Chukchi Sea reaches to the shelf bottom and that the water is enriched by nutrients regenerated from the sediments. The nutrient maximum has been followed from the Canadian Basin into the western Eurasian Basin and Fram Strait as well as into the Canadian Arctic Archipelago. In both basins the densest part of the halocline has a salinity of 34.2–34.4 PSU. This layer is initially formed out of the Fram Strait branch north of Svalbard through the melting of sea ice on top of the warm AW. It is advected eastward with the boundary current and becomes homogenized to deeper than 100 m by winter convection and in summer it is capped by less-saline melt water. The southern Nansen Basin therefore does not exhibit the common PML-halocline structure but just a deep winter mixed layer reaching to the thermocline. North of the Laptev Sea this winter mixed layer, because of the outflow of low salinity shelf water, becomes part of the halocline, and as it becomes isolated from the winter convection, its temperature increases by mixing with the underlying Atlantic Layer (**Figures 4A** and **5A**). Over most of the Eurasian Basin, the PSW consists only of this dense part of the halocline and the PML.

Below the halocline both temperature and salinity increase with depth and a layer with temperatures above 0°C is encountered. This layer was observed by Nansen on Fram, and identified by him as deriving from the AW entering through Fram Strait. In the Eurasian Basin the Atlantic Layer lies between 200 m and 700 m and has a maximum temperature of 1–3°C. In the Canadian Basin the maximum temperature is normally below 1°C, in the Canada Basin even below 0.5°C, and is found between 300 m and 800 m (**Figure 5**). In the interior of the basins the strong stability limits the upward heat loss from the Atlantic Layer, and its lower temperatures and salinities compared with the AW in Fram Strait reflect the mixing with the Barents Sea branch water, as well as the incorporation of cold plumes into the boundary current at the slope.

The intermediate water colder than 0°C is here called upper Polar Deep Water (uPDW). Its temperature decreases with depth from 0°C to about -0.5°C, while its salinity increases from 34.85 PSU to above 34.9 PSU. The uPDW is warmer and more saline in the Canadian Basin than in the Eurasian Basin. This is due to different formation processes. In the Eurasian Basin the uPDW range is dominated by the inflow of the Barents Sea branch, which creates a salinity minimum in the water column. In the Canadian Basin the uPDW in the entering boundary current is transformed by slope convection. Saline plumes sink through the Atlantic Layer, entraining warmer water and redistributing it downward into the uPDW. This creates the stable potential temperature–salinity (Θ–S) signature characteristic of the intermediate waters in the Canadian Basin (**Figure 5**). Below the salinity minimum the slope of the Θ–S curves in the Eurasian Basin is similar to that in the Canadian Basin.

The characteristics of the two basins continue to diverge in the deep waters. In the Eurasian Basin there is a 1000 m thick layer with almost constant salinity and decreasing temperature, while the temperature decrease in the Canadian Basin is much smaller (only to -0.55°C), but the salinity increases to above 34.95 PSU in the same depth range. The Eurasian Basin deep Water (EBDW) thus becomes colder ($\Theta < -0.9$°C) and less saline (34.92–34.93 PSU) than the Canadian Basin Deep Water (CBDW). This agrees with a strong input from slope convection in the Canadian Basin water column, while in the Eurasian Basin the effects of the Barents Sea and Fram Strait inflows are still recognized at these levels. Below the sill depth of the Lomonosove Ridge, approximately coinciding with the $\sigma_{1.5} = 35.142$ isopycnal, the waters of the two basins thus differ considerably (**Figure 5**).

Below 2000 m the salinity of the EBDW starts to increase and reaches a maximum above 34.94 PSU at the bottom. The temperature remains almost constant, but a small minimum is found about 800 m above the bottom (**Figure 5**). The increasing salinity indicates that deepest layers of the Eurasian Basin are renewed by slope convection. The temperature is lower than in the Canadian Basin because the water from the most probable sources, the deepest part of the St Anna inflow and the shelf area around Severnaya Zemlya, either starts entraining first at 1000 m, or sinks through, and entrains, cold Barents Sea branch water, not the warmer Fram Strait branch present at the slope in the Canadian Basin. Although the dense shelf water triggers the convection, it supplies only a small fraction (<5%) to the deep waters. The main contribution is waters from shallower levels entrained into the plumes.

The salinity of the CBDW remains constant in the deepest 1000 m while its temperature decreases slightly (**Figure 5**). A possible explanation is that colder water from the Amundsen Basin passes into the Makarov Basin through gaps in the central part of the Lomonosov Ridge. If it is displaced downward, its density relative to the Canadian Basin water column will increase because of the larger compressibility of colder water (the thermobaric effect) and it continues down to the bottom. This additional source for the deep waters in the Canadian Basin, which does not entrain the warmer AW of the boundary current, dominates the deepest layers of the Canadian Basin.

The CBDW is warmer than the EBDW and remains, because of the thermobaric effect, at sill depth as it enters the Eurasian Basin. An intermediate salinity maximum around 1700 m in the Amundsen Basin (**Figure 5**), strongest close to Greenland, agrees with a return flow from the Canadian to the Eurasian Basin along the Greenland continental slope, which partly becomes deflected into the Amundsen Basin at the Morris Jesup Plateau (**Figure 6**). Weaker, intermittent, spillover between the two basins is likely to occur along the entire Lomonosov Ridge.

The difference between the deep waters of the two basins have commonly been explained by assuming that no deep water is formed in the Arctic Ocean but is supplied from the Greenland Sea through Fram Strait. The higher temperatures in the Canadian Basin would then be due to the blocking of the coldest and densest water by the Lomonosov Ridge. The existence of the Lomonosov Ridge was actually deduced in this manner. The examination of the deep water properties shows that this is not the main cause. An inflow of colder, less saline, intermediate and deep waters through Fram Strait does take place (**Figure 5B**). It contributes to, but it does not determine, the characteristics of the Arctic Ocean deep waters, especially not at the deepest levels (>2600 m), and it cannot explain the higher temperatures and salinities in the Canadian Basin.

Outflows

The Arctic Ocean waters exit primarily through the Canadian Arctic Archipelago and in the East Greenland Current through Fram Strait, while the Barents Sea is less important. The upper waters in the northern Barents Sea have a salinity of ~ 34.5 PSU and derive from low-salinity water and ice entering from east and north. The net transport to the Norwegian Sea is small, because the West Spitsbergen Current

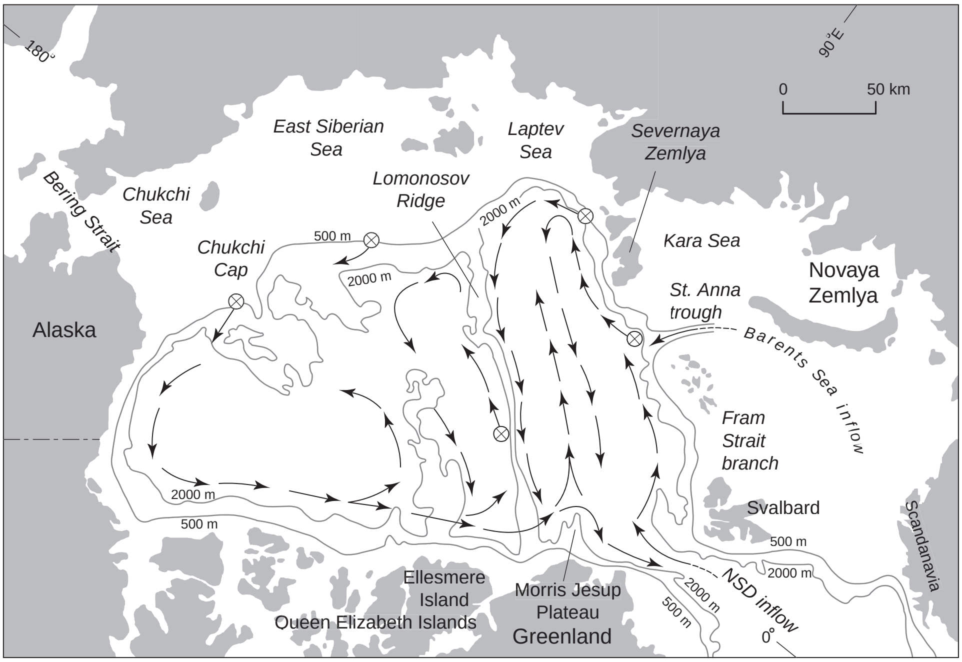

Figure 6 Schematics indicating the circulation of the deeper layers and possible source areas for the deep waters (crosses). (Adapted from Jones EP, Rudels B and Anderson LG (1995). Deep waters of the Arctic Ocean: Origins and Circulation. *Deep-Sea Research I* 42: 737–760.)

carries most of the low-salinity waters back to the Arctic Ocean.

The outflow through the Canadian Arctic Archipelago consists mainly of low-salinity ($32.5 < S < 33$ PSU) upper waters, mostly derived from the Bering Strait inflow. The properties of the bottom water in the Baffin Bay ($\Theta \sim -0.5^{\circ}$C, $S \sim 34.45$ PSU) show that an outflow of water from the upper part of Arctic Ocean thermocline also must occur. The total transport is 1–1.5 Sv.

Fram Strait is the only deep passage, and the East Greenland Current transports, in addition to 1 Sv of PSW, intermediate waters as well as CBDW and EBDW. The uPDW makes the -0.5 to 0.0°C temperature range more prominent at the western slope of Fram Strait. The CBDW show up as a salinity maximum at the western slope at about 1800 m, and still deeper the colder and saline EBDW is seen (**Figure 7**). The outflow of intermediate waters is about 2–3 Sv, that of deep waters 0.5–1 Sv.

Variability

Observations in the 1990s have revealed large changes in the Arctic Ocean water masses. The AW in the boundary current has been almost 2 degrees warmer than the climatological mean. This warm pulse has been followed into the different basins, and along the different circulation loops. At the same time the intermediate water below the temperature maximum has become less saline. In the eastern Amundsen Basin the salinity of the PML has increased, and in 1995–1996 a deep, saline (34.3 PSU) mixed layer extended down to the thermocline. In the Canada Basin, by contrast, the salinity of the PML has become lower. The extent of the Beaufort gyre has been reduced, and the Pacific Water has been forced closer to the American continent. The boundary between Pacific and Atlantic derived upper waters appears to have shifted from a previous position above the Lomonosov Ridge to above the Mendeleyev Ridge.

The average thickness of the ice cover decreased from 3.1 to 1.8 m between the 1970s and the 1990s. The thinning is most prominent in the central and eastern part of the Arctic Ocean. The larger fresh water content of the PML in the Canada Basin could then be related to melting of sea ice, while the higher salinities in the Amundsen Basin suggest that other causes, such as increased ice export, also may be important. Higher air temperatures would lead to less ice formation and a thinner ice cover. The exceptionally warm Atlantic Layer and the absence of a halocline in the PSW could allow for a larger heat flux from the Atlantic Layer to the surface water that also would reduce ice formation.

The hydrographic structure in Fram Strait has changed. In 1984 the recirculating AW extended to the Greenland slope, while in 1997 it was more concentrated to the centre of the strait, leaving a wider passage for the intermediate waters exiting the Arctic Ocean (**Figure 7**). This suggests that the recirculation of the West Spitsbergen Current in Fram Strait has weakened and that more AW enters, and follows, the different loops in the Arctic Ocean basins.

The North Atlantic Oscillation, a large-scale decadal mid-latitude phenomena, has been invoked to explain these changes. Its state is commonly described by the NAO index, the difference in winter surface air pressure between the Azores high-pressure and the Icelandic low-pressure systems. A high index implies strong cyclonic activity, and warmer air over the Nordic Seas. Less cooling occurs, and the AW entering through Fram Strait will be warmer. A larger precipitation over Scandinavia decreases the salinity of the Norwegian Coastal

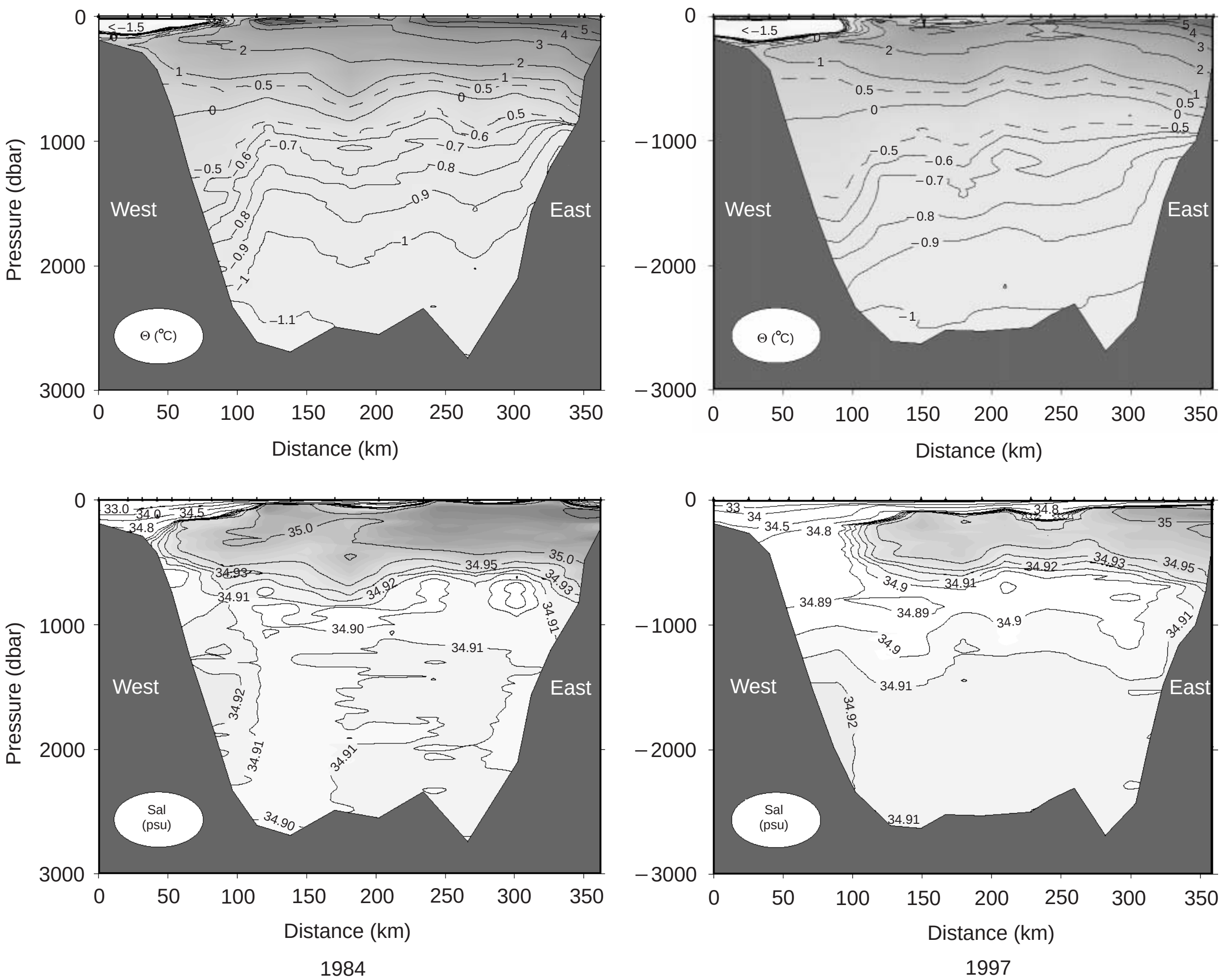

Figure 7 Potential temperature and salinity sections along the sill at 79°N in Fram Strait taken by R.V. Lance in August 1984 and in August 1997. The structure is determined by the two boundary currents, the West Spitsbergen Current in the east and the East Greenland Current to the west. The upper 500 m are dominated by the inflowing and recirculating AW of the West Spitsbergen Current. The cold, low salinity PSW is located over the Greenland shelf and slope. The Atlantic Layer and the uPDW in the East Greenland Current are identified by the increased separation of the -0.5 and $+0.5$ isotherms at the Greenland slope. The CBDW is seen as a salinity maximum at 1800 m, which extends down to the EBDW. The northward flow of the colder and less saline Arctic Intermediate Water (AIW) and Nordic Sea Deep Water (NSD) occurs mainly at the Svalbard slope, where the isotherms rise and the -0.5 to $+0.5$ interval is all but absent. A second core of NSD is present just east of the Greenland slope, most conspicuous in 1984. The AIW is the most likely source for the salinity minimum extending across the strait at 800 m. Large changes have occurred between 1984 and 1997. The AW is less saline, and the recirculation of the West Spitsbergen Current does not extend to the Greenland slope, leaving a broader passage for the intermediate waters exiting the Arctic Ocean. The temperature and salinity of the deeper layers have increased, and no bottom water colder than −1.1°C and with salinity below 34.9 PSU is seen in 1997. This implies a stronger presence of Arctic Ocean deep waters. The salinity minimum at 800 m has become more prominent, suggesting a stronger input of AIW while the northward flow of NSD has weakened. This is consistent with the observations that the Greenland Sea at present produces AIW rather than Greenland Sea Deep Water.

Current and leads to a lower salinity in the Barents Sea branch, which shows up in the Arctic Ocean as lower salinity at intermediate depth. Cyclones penetrating into the Arctic weaken the high-pressure cell over the Beaufort Sea, forcing it closer to the American continent. They could also make the low-salinity shelf water in the Laptev Sea enter the East Siberian Sea and the Canadian Basin rather than the Amundsen Basin. The salinity in the Canadian Basin would then become lower, while, in the absence of a PML, a deep (>100 m), saline (34.3 PSU) mixed layer characterizes the Amundsen Basin as well as the Nansen Basin.

Such drastic changes may frequently occur in the Arctic Ocean but have, until now, remained undetected due to lack of observations. The NAO, however can be followed over 100 years. In the 1990s it was at its highest recorded value and exhibited larger variability than ever before.

Climatic Significance

The stable stratification of the upper waters allows for the formation, and existence, of a perennial ice cover, which in summer, because of its higher albedo, reduces the absorption of incoming solar radiation. The ice cover insulates the ocean and lowers the heat loss to atmosphere and space. The stratification is created by runoff and by the inflow of low-salinity water from the Pacific Ocean. A low-salinity surface layer would, however, also form without these inputs because of the separator process discussed above. On the shelves the entire water column is cooled to freezing temperature in winter, leading to ice formation and brine rejection. In summer the ice melts and creates a low-salinity surface layer that, when exported to the basins, provides the stability necessary for ice formation and the existence of an ice cover also in the interior of the Arctic Ocean. The salinity of the PML, formed by this separation process, would be higher than today, and the ice cover thinner and more vulnerable to heat flux from the underlying Atlantic Layer.

The Arctic Ocean contributes the Meridional Overturning Circulation (MOC) by providing dense waters to the Greenland-Scotland overflow. The intermediate waters cross the sill but the denser CBDW and EBDW that remain in the Arctic Mediterranean also contribute by displacing less-dense water in the Nordic Seas to shallower levels. A total of ~ 3 Sv of the overflow is supplied by the Arctic Ocean. Reduced heat loss in the Arctic as well as a larger input of fresh water to the Arctic Ocean would decrease the production of these waters, and more low-salinity, upper waters would be exported through Fram Strait. This could, if an excessive amount enters the Greenland and Iceland Seas, influence, and temporarily reduce, the deep water formation in these seas. As the low-salinity water continues around Greenland as the West Greenland Current, it could, together with a less-saline outflow through the Canadian Arctic Archipelago, also affect the deep convection in the Labrador Sea. Such changes would weaken the MOC, reducing the compensating inflow of warm surface water to the Arctic Mediterranean across the Greenland-Scotland Ridge, and influence the climate of the Arctic and of north-western Europe.

See also

Bottom Water Formation. Double-diffusive Convection. Freshwater Transport and Climate. Heat Transport and Climate. Intrusions. Meddies and Sub-surface Eddies. Non-rotating Gravity Currents. North Atlantic Oscillation (NAO). Ocean Circulation. Open Ocean Convection. Overflows and Cascades. Polynyas. Rotating Gravity Currents. Sea Ice: Overview. Thermohaline Circulation. Upper Ocean Heat and Freshwater Budgets. Upper Ocean Mixing Processes. Water Types and Water Masses.

Further Reading

Coachman LK and Aagaard K (1974) Physical oceanography of the Arctic and Sub-Arctic Seas. In: Herman Y (ed) *Marine Geology and Oceanography of the Arctic Ocean*, pp. 1–72. New York: Springer.

Johannesen OM, Muench RD and Overland JE (eds) (1994) *The Polar Oceans and Their Role in Shaping the Global Environment*, AGU Geophysical Monographs 85. Washington, D.C.: American Geophysical Union.

Leppäranta M (ed) (1998) *Physics of Ice-covered Seas*. Helsinki: Helsinki University Press.

Lewis EL (ed) (2000) *The Freshwater Budget of the Arctic Ocean*. Dordrecht: Kluwer Academic Publishers.

Nansen F (1902) Oceanography of the North Polar Basin. *The Norwegian North Polar Expedition 1893–96. Scientific Results III (9)*, Christiania: Jacob Dybwad.

Smith WO Jr (ed) (1990) *Polar Oceanography*, part A: *Physical Sciences*. San Diego: Academic Press.

Smith WO Jr and Grebmeier JM (eds) (1995) *Arctic Oceanography, Marginal Ice Zones and Continental Shelves*, Coastal and Estuarine Studies, Volume 49. Washington DC: American Geophysical Union.

Untersteiner N (ed) (1986) *The Geophysics of Sea Ice*. New York: Plenum Press.

Wadhams P, Gascard J-C and Miller L (eds) (1999) Topical studies in oceanography: The European Subpolar Ocean Programme: ESOP. *Deep-Sea Research II* 46: 1011–1530.

Wheeler PA (ed) (1997) Topical studies in oceanography: 1994 Arctic Ocean Section. *Deep-Sea Research II* 44: 1483–1758.