



The circulation and water mass (trans)formations in the Arctic Mediterranean Sea and their implications for the deep ocean circulation: a narrative

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Abstract. The Arctic Mediterranean is rapidly changing, a statement that is often made, but the follow-up statements: changing from what and towards what are often omitted. Hence its role in the global climate system, particularly regarding the Atlantic Meridional Overturning Circulations remains poorly constrained. This review of the oceanography of the Arctic Mediterranean Sea develops an unified perspective of how interacting components of the system evolve in space and time, and the processes that determine their evolution. To set the stage a succinct overview of the geographic setting and early explorations is given. We then follow the pathways of the principal water masses to describe inflows, inter-basin circulations, water mass transformations, and outflows of heat and salt to the bordering subpolar gyre and the global ocean. The fundamental connection to the global ocean, the Atlantic water, is traced along its route into and through the Arctic Mediterranean. Its transformations, driven by cooling, and by the freezing and melting of sea ice, lead to the creation of both denser and less dense waters that form and maintain the water column structures within the Arctic Mediterranean; the so-called double estuary. A second advective component to the Arctic, the low salinity water carried by the Pacific inflow, is concentrated into the Amerasian Basin and acts there to further isolate the denser waters derived from the Atlantic inflow. Hence, the waters return to the North Atlantic either as dense overflows or buoyant outflows. The water masses within diverse regions of the Arctic Mediterranean have changed over the past few decades, and this influences their exchanges with the world ocean east and

west of Greenland. That the water mass transformations in the Arctic Mediterranean take place beyond a ridge system that allows for the build-up of significant density differences that, through entrainment, can increase the impact of the Arctic Mediterranean on global overturning circulation.

1 Introduction

The mean surface temperature of the earth is rising and how this will influence the circulation of the atmosphere and in the oceans is largely unknown. The observed temperature increase is strongest in the high northern latitudes, which are primarily covered by ocean and one question then becomes: Does the presence of an ocean accelerate or mitigate the observed changes? To address this and related questions we review in Sect. 2 the global radiation balance and how the high north responds to the radiative and mechanical forcing generated by the sun and by the atmosphere. The focus is on the Arctic Mediterranean Sea, which comprises the Arctic Ocean and the Nordic Seas. We also describe the physical and geographical settings and the main bathymetric features of the Arctic Mediterranean and its connections to and communications with lower latitudes in the ocean and in the atmosphere.

Section 3 describes how our knowledge of the Arctic has evolved during the 500 years that have elapsed since the exploration of the high northern latitudes started. This section also reviews how the knowledge of the circulation and the water mass (trans)formations taking place in the Arctic

Mediterranean gradually evolved during the about 150 years that oceanographic observations and studies have been conducted in the Arctic Mediterranean. It also discusses the early views of the deep circulation in the world ocean and finally brings up the establishing of the Committee for the Exploration of the Sea (ICES) in 1899 and discusses its importance for the improvement of measurement techniques and theoretical understanding of the ocean circulation.

Section 4 examines the different inflows to the Arctic Mediterranean, the Pacific inflow through Bering Strait and the North Atlantic circulation from the subtropical gyre to the subpolar gyre and into the Nordic Seas. The next sections concentrate on specific features, areas and processes that are presently active and may become more (or less) important in a future warmer climate. Section 5 focusses on sea ice, its physical properties and how its growth and melting influence the underlying water column. Section 6 describes the two inflows from the Nordic Seas to the Arctic Ocean, through Fram Strait and over the Barents Sea, and the interactions (freezing and melting) between Atlantic water and sea ice. Section 7 concentrates on the transformations of the Atlantic water as it moves eastward along the two inflow paths and on the subsequent meeting and remerging of the two inflow branches at the continental slope north of the Kara Sea, Severnaya Zemlya and the Laptev Sea. In this section a wind driven model proposed by Nøst and Isachsen (2003) is also presented.

Eastward of the Laptev Sea the influence of the Atlantic water diminishes and the circulation becomes dominated by freshwater input, both as continental runoff from the Eurasian rivers and as direct inflow of low salinity Pacific water through the shallow Bering Strait (Sect. 8). The upper ocean circulation and mixing are determined by wind and by the seasonal changes in thermohaline forcing, while the deeper (Atlantic) layers become isolated and the observed changes appear due to internal mixing processes such as thermohaline intrusions and double-diffusively active interfaces. In the Transpolar Drift the Siberian branch transports runoff and sea ice from the Laptev Sea, and the anticyclonic Beaufort Gyre brings additional low salinity water towards Greenland. The convergence of these two streams leads to exchanges between the two gyres; the cyclonic gyre in the Eurasian Basin and the anticyclonic gyre in the Amerasian Basin. In Sect. 8 the differences between the deep waters in the Amerasian and Eurasian basins and their causes are also described and discussed.

In Sect. 9 the water masses exchanged through Fram Strait are described from two hydrographic sections, one north of the strait and one along the sill, both taken by FS *Polarstern* in 2004. Different water masses are identified and the transports of volume, freshwater and heat through the strait are determined from geostrophy combined with inverse methods and a discussion about the reasonableness of the obtained mass and freshwater transports is provided.

The Nordic Seas, especially the Greenland Sea, are characterized by weak stability and are areas of deep convection and deep water renewal, and the convection in the Greenland Sea is discussed in Sect. 10. The water column in the Greenland Sea has exhibited large changes during the last 50 years, indicating that the nature of the active convection processes has changed (compare, for example Carmack and Aagaard, 1973 and Brakstad et al., 2019). This is perhaps the area, where the effects of a warming climate on the water column have been the strongest.

The three overflows of dense water across the Greenland–Scotland Ridge, through the Faroe Bank Channel, across the Iceland Faroe Ridge, and through Denmark Strait, are examined in Sect. 11. The focus is on the Denmark Strait overflow, its possible sources, and the entrainment as the overflow plumes descend into the deep North Atlantic.

The outflows in the upper layer that take place through Fram Strait in the East Greenland Current and through the channels in the Canadian Arctic Archipelago to Baffin Bay and continuing through Davis Strait to the Labrador Sea are discussed in Sect. 12. As the East Greenland Current passes Cape Farewell it turns north as the West Greenland Current and enters Baffin Bay from the south. The transports are modelled as geostrophic flows through control sections with the coast at the righthand side (Werenskiold, 1935). The Canadian Arctic Archipelago and Baffin Bay are considered as a connected and interacting system, and the transports determined from geostrophy are combined with estimates of the surface heat loss and ice formation that determine the upper layer characteristics in Baffin Bay. Estimates of the freshwater storage in the Arctic Ocean are also given and the effects of increased freshwater input from the Greenland icecap are discussed.

The exchanges between the Arctic Mediterranean and the subpolar gyre and the Labrador and Irminger seas are described in Sect. 13, where also the impacts of water mass transformations taking place in a semi-isolated sea may have on the overturning circulation are examined. The final Sect. 14 discusses the global connections and interactions and how they influence and are influenced by the Arctic Mediterranean Sea.

The present review intends to give a comprehensive picture of the Arctic Mediterranean Sea, describing the processes and the water mass transformations in the different parts of the area and how these parts are connected by the circulation of the different waters. The narrative follows two tracks. The first is historical and describes how our knowledge of the geography, the hydrography and the ocean circulation has evolved through time. This approach continues to about 1980 and encompasses the knowledge provided by early summaries of the circulation in the Nordic Seas (Helland-Hansen and Nansen, 1909), the connections between the North Atlantic and the Nordic Seas (Helland-Hansen, 1912; Nansen, 1912, 1913) and the early knowledge of the water masses and circulation in the Arctic Ocean

(Nansen, 1902, 1906; Timofeyev, 1960; Coachman and Aagaard, 1974).

After this focus shifts, and the narrative follows the Atlantic water as it enters the Arctic Ocean through Fram Strait and over the Barents Sea, and how it becomes transformed by the processes within the Arctic Ocean, especially by interactions with sea ice and with the freshwater input from the continents and from the Pacific inflow. The Atlantic water is transformed into both less dense water (outflow) and denser water (overflow) within the Arctic Ocean and additional overflow water is formed in the Greenland Sea. The transformed waters eventually return to the subpolar gyre and the North Atlantic.

The review is primarily based on observations and their interpretations. The figures shown thus display mainly temperature and salinity profiles and TS diagrams and vertical temperature and salinity sections. This to show what the interpretations and discussions are based upon. The only schematics are presented in the last figure (Fig. 24).

There are many reviews addressing the Arctic Mediterranean Sea, the Nordic sea and the Arctic Ocean. Coachman and Aagaard (1974) give a foundational synthesis of Arctic and subarctic hydrography, circulation, exchange passages, and sea-ice conditions based on the observational knowledge available in the early 1970s. It establishes the Arctic Ocean as a strongly stratified, relatively isolated basin connected to the Pacific and Atlantic through a few narrow gateways. It describes the importance of Bering Strait inflow, the Atlantic inflow via the Nordic Seas and Fram Strait, and the freshwater and ice export to the North Atlantic.

An early modern synthesis of the physical oceanography of the Nordic (or GIN) sea is provided by Hopkins (1991). It is the major physical-oceanography review of the Greenland, Iceland, and Norwegian Seas for the 1970s to the mid-1980s. It describes the basin hydrography, fronts, gyre circulation, the Atlantic Water inflow, the Arctic-origin outflow, deep convection, and sea-ice distribution and explains how the Nordic Seas transform northward-flowing Atlantic waters through cooling, mixing, and densification.

A summary of the pre-IPY (International Polar Year) 2007–2009 knowledge of Arctic water masses and pathways is given in Rudels (2015). It explains the layered structure: the polar mixed layer, the halocline, the Atlantic Water, and deep and bottom waters, and highlights dense-water formation on shelves, boundary current circulation and exchanges with the Nordic Seas and Pacific.

The reviews most akin to the present one are written by Rudels (2012, 2019), Timmermans and Marshall (2020), and Rudels and Carmack (2022). Rudels (2019) gives a concentrated description of the Arctic Ocean circulation and water mass transformations, while Rudels (2012) covers more of the history of Arctic research. Timmermans and Marshall (2020) describe the dynamically distinct Arctic circulation: wind forcing, buoyancy forcing, topographic steering, eddies, and boundary currents. They emphasize how freshwater

storage, sea-ice loss, and changing Atlantic or Pacific inflows may reshape stratification and circulation. Rudels and Carmack (2022) discuss the Atlantic and Pacific inflow branches, halocline formation, shelf contributions, and deep-water renewal, and stresses the controlling role of ridges, shelves, and narrow gateways in partitioning water masses among the Arctic basins.

Bluhm et al. (2015) compare the Eurasian and Amerasian basins, including their water-mass structures, circulation, ventilation, and connectivity and show how the warmer, saltier Atlantic influence is more direct in the Eurasian Basin, while the Canada Basin is more strongly influenced by freshwater accumulation and Pacific-origin waters. The review by Wassmann et al. (2020) and book edited by Wassmann (2024) also connects basin-scale physical contrasts to ecosystems and biogeochemistry.

Bacon et al. (2022) concentrates on the Arctic Ocean boundary exchanges and describes the observational estimates of volume, heat, freshwater, and property exchanges through the Arctic gateways. It covers Fram Strait and the Barents Sea Opening in particular, making it directly relevant to the Arctic ocean–Nordic Seas exchange. It emphasizes that gateway fluxes, surface forcing, and storage changes together control Arctic water-mass transformation.

We have perhaps treated sea ice processes and the observed changes in the ice cover rather cavalierly, and we here refer the reader to Meier et al. (2014), which is a broad review of observed losses in Arctic sea-ice extent, thickness, age, and multiyear-ice coverage. It attributes the shift to coupled atmospheric, oceanic, and ice–albedo feedbacks, with summer melt increasingly sensitive to preconditioning. It also discusses implications for Arctic communities, ecosystems, shipping, resource activity, and weather and climate connections

The book by Rudels (2021) contains much of the topics discussed in our review, but it is a textbook, not a review, and contains more detailed descriptions of the circulation and the active processes than are necessary to include in a review article. The book by Mills (2009) provides a very readable account of the history of the ocean sciences.

This review is, admittedly, focused on Western science, as based on the scientific method, and gives little recognition to indigenous contributions to exploration and ecological knowledge. There are two main reasons for this. First, it is beyond the scope of this review, and the authors lack the necessary expertise to do this complex subject justice. Second, the focus of this review is on the circulation and transformations of waters in the deep and offshore Arctic Mediterranean, and thus remote from the Arctic coastal regions where indigenous peoples travel, hunt and live. Regardless, there is much to learn from indigenous thinking as our global society addresses an uncertain future (cf. Berkes and Jolly, 2001; Ford et al., 2010; Falardeau and Bennett, 2019; Falardeau et al., 2019; Huntington, 2011; Huntington et al., 2021). Indigenous knowledge resides in the

holder's authority and is experiential and holistic and place-based (Gearheard et al., 2006; Whiteman and Cooper, 2011; Fienup-Riordan and Carmack, 2011). Indigenous knowledge is not a single, uniform body of information, but is learned, practiced, and transmitted differently among distinct Arctic Indigenous Peoples; for reviews, see Eerkes-Medrano and Huntington (2021). Despite the above distinctions, there are key areas where complementarity exists (cf. Carmack and Macdonald, 2008). Intimate connections and dependence of snow and ice conditions can provide an early-warning system for climate and ecological change; one clear example is in documenting local observations based intergenerational knowledge and travel/hunting practice as records of environmental change (Krupnik and Jolly, 2002; Carmack, 2021). That the two-ways of knowing are jointly beneficial was demonstrated by comparison of remote sensing of sea ice and Inuit observations (Meier et al., 2006). Fienup-Riordan and Carmack showed the complementarity of broad spatial records (wind-generated ocean swell) and fine-scale knowledge of ice quality and currents (safety of travel over ice in coastal locations). We emphasize that the depth, breadth and sophistication of Indigenous Knowledge of what is Arctic equals that of Western science, and future avenues based on appropriate co-generation of knowledge offer a rich field of inquiry.

2 The North Polar Setting

The earth receives its energy as shortwave radiation from the sun and its rapid rotation limits the diurnal temperature differences between the sunlit and the dark hemispheres. Large temperature differences arise, however, because of the spherical shape of the earth. The zenith angle of the sun increases with latitude and higher latitudes receive less solar radiation per unit area. The energy received on the luminated hemisphere is radiated back to space as longwave radiation from the entire earth surface. To zero order the energy input and output are globally but not locally balanced and the net heat input at lower latitudes is transported by the atmosphere and the oceans to higher latitudes, where it is radiated back to space. The presence of an atmosphere that contains greenhouse gases that absorb and reradiate longwave radiation leads to a higher mean temperature of the earth than if no atmosphere were present.

The earth's rotation axis makes an angle of 23.5° to the ecliptic. This leads to large seasonal variations in incoming solar radiation, especially north and south of the polar circles ($> 66.5^\circ$). In summer the incoming daily radiation at the top of the atmosphere is larger at the poles than at the equator (e.g. Fritz, 1951), while in winter the sun is below the horizon and the long, dark polar night reigns. The contrast between the two polar regions is, however, great. In the south the entire ice covered Antarctic continent is in direct contact with the circumpolar Southern Ocean, while in the north there is a

confined ocean, the Arctic Mediterranean Sea, with restricted communications to the world ocean and with its northernmost part, the Arctic Ocean, partly covered by perennial sea ice.

The atmospheric poleward heat transport is carried by three distinct circulation cells, the direct Hadley cell where warm air rises at the equator, cools and then sinks back to the surface at 20 to 25° N(S). Some of the sinking air returns towards the equator, forming the trade winds, while the rest is transported poleward by the westerlies in the eddy dominated Ferrell cell. Over the polar areas, in the direct Polar cell, the air is cooled, sinks and returns equatorward, forcing the warmer air of the Ferrel cell to rise.

At low latitudes the poleward oceanic heat transport is larger than the corresponding atmospheric transport, but the transports become equal around 30° N(S) (Wunsch, 2005). At higher latitudes the ocean transport diminishes and beyond 65° only the heat transport in the North Atlantic is different from zero. The Atlantic also differs from the other oceans in that it transports heat across the equator from the southern to the Northern Hemisphere. The warm surface water ultimately comes from the Indian Ocean and is carried by the South Atlantic circulation to and across the equator.

The atmospheric circulation transports not only heat but also water vapor from lower to higher latitudes. In the south this builds up and sustains the glaciers on the Antarctic continent and the freshwater returns as calving icebergs to the Southern Ocean. In the north the water condenses, releasing latent heat and then precipitates, partly over the Arctic Mediterranean, but primarily on the continents surrounding the Arctic Ocean. The drainage area is twice as large as the surface area of the Arctic Ocean and provides a disproportionate freshwater input, 10 % of the entire global continental runoff, to the Arctic Ocean. The created low salinity upper layer leads to strong stability that limits the interaction with the deeper part of the water column.

In summer the net heat input from the sun is larger than the net outgoing longwave radiation and the radiation balance is positive (more heat received than lost). However, the outgoing longwave radiation and the downward reradiated longwave radiation are both individually larger than the downward shortwave solar radiation. During the long winters with no local energy input from the sun the heat advected from lower latitudes cannot balance the radiative heat loss to space and the surface temperature decreases drastically.

The surface air temperature in the Arctic is above the melting temperature of sea water in summer but much below in winter. This leads to phase changes. Sea ice is formed in winter and melts again in summer. The latent heat released by freezing becomes a local heat source that somewhat increases the surface temperature in winter. In summer the returning solar radiation raises the surface temperature and eventually the ice cover starts to melt. The melting ice acts as a thermostat that keeps the surface water temperature close to the melting temperature as long as ice is present. Ice formed and

melted does not contribute any net heat to the Arctic Ocean, but the melting in summer stabilizes the water column and isolates the deeper layers from the surface forcing in summer even more than in winter. The ice not melted in summer is eventually exported by the ocean currents southward and becomes a heat source to the Arctic.

The Arctic Mediterranean Sea is a part of the North Atlantic and comprises the Arctic Ocean and the Nordic Seas; the Greenland, Iceland and Norwegian seas, that form an anteroom to the larger Arctic Ocean. The Nordic Seas communicate to the south with the North Atlantic across the Greenland–Scotland Ridge and are connected to the Arctic Ocean through Fram Strait and over the Barents Sea in the north. There is also a direct connection between the Arctic Ocean and the North Atlantic through the narrow straits in the Canadian Arctic Archipelago and then via Baffin Bay and Davis Strait to the Labrador Sea. The Arctic Ocean is connected to the North Pacific through Bering Strait. It is thus not a bay but a convoluted channel between the stratified North Pacific and the convectively ventilated North Atlantic.

The Arctic Ocean is the smallest of the Earth's oceans ($9.5 \times 10^{12} \text{ m}^2$), if it could be called an ocean, comprising only about 3 % of the area and less than 1 % of the volume of the global ocean. More than half of the area of the Arctic Ocean consists of broad, shallow shelf seas mainly located north of the Eurasian continent. Farthest to the west is the Barents Sea, separated from the Kara Sea by Novaya Zemlya and Franz Josef Land. The Kara Sea connects eastward to the Laptev Sea through Vilkitsky Strait between Siberia and Severnaya Zemlya. The New Siberian Islands separate the Laptev Sea from the East Siberian Sea, which communicates through Long Sound between the Siberian mainland and Wrangel Island with the Chukchi Sea located north of Bering Strait. Eastward of Alaska the coastline is different. Instead of shallow shelf seas and low lying coasts, it forms the mountainous Canadian Arctic Archipelago, comprising several large islands creating a maze of channels. The most important passages are Lancaster Sound, Jones Sound, and the Nares Strait between Ellesmere Island and Greenland that all connect the Arctic Ocean to Baffin Bay (Fig. 1).

The deeper part of the Arctic Ocean consists of two major basins, the Amerasian Basin and the Eurasian Basin, separated by the narrow Lomonosov Ridge with a mean sill depth around 1600 m. The Amerasian Basin is divided by the Alpha and Mendeleev ridges into the large Canada Basin and the smaller Makarov Basin, both with maximum depth of the abyssal plains around 4000 m. The Eurasian Basin also consists of two basins, the about 4000 m deep Nansen Basin between the Eurasian continental slope and the Gakkel Ridge and the deeper, 4500 m, Amundsen Basin extending from the Gakkel Ridge to the Lomonosov Ridge. The connection to the Nordic Seas through Fram Strait is deep, sill depth 2600 m. The Greenland Sea comprises two basins, the Greenland Basin with maximum depth about 3500 m and the northerly, smaller and shallower Boreas Basin. The Norwe-

gian Sea is separated into the Norwegian Basin and the Lofoten Basin both about 3000 m deep, while the Iceland Sea is shallower, less than 2000 m deep.

Three deeper passages through the Greenland–Scotland Ridge connect the Nordic Seas to the North Atlantic, the 640 m deep Denmark Strait between Greenland and Iceland, the 400 to 500 m deep ridge between Iceland and the Faroes, and the channel between the Faroes and Scotland, ending in the more than 800 m deep Faroe Bank Channel. West of Greenland the channels from the Arctic Ocean to Baffin Bay are narrow and shallow. Nares Strait has a sill depth of 220 m and Lancaster Sound and Jones Sound have sill depths between 100 and 150 m. Baffin Bay is deep, about 2600 m, and the sill depth in Davis Strait to the Labrador Sea is 630 m. Bering Strait between the North Pacific to the Arctic Ocean is narrow, 85 km, and shallow, sill depth 53 m.

3 Exploration

3.1 voyages of discovery

It was after Bartolomeu Diaz rounded the Cape of Good Hope in 1488 and the discovery of a new world by Christopher Columbus in 1492 that the exploration of the northern waters of the Atlantic Ocean began in earnest. The papal meeting in Tordesillas in 1494 had divided the world into two spheres of interest. The one east of 49° W should be explored and exploited by Portugal and the one west of that longitude by Spain. The newly discovered American continents constituted a barrier for a direct sailing to Cathay and the treaty excluded other nations such as England, France and Holland from using the southern routes to India and China. This forced them to explore the east coast of the North American continent to find a passage westward from the Atlantic. The Genua born John Cabot made the first English sailing northward along the American coast already in 1497 but found no passage. This was 16 years before Vasco de Balboa first looked upon the Southern (Pacific) Ocean from the Mexican western shore and 25 years before Magellan sailed through the strait named after him and discovered the southern passage to the Pacific Ocean and to China and India.

The efforts to find a northern route between the Atlantic and Pacific continued, however, and the possibilities to find both a Northwest Passage through or north of North America and a Northeast Passage north of the Eurasian continent were explored but with no success. The expeditions were stopped by ice and many ended with the loss of ship and crew. It would take 300 years before both passages had been navigated by a single vessel, the Northeast Passage by *Vega* 1878–1880 and the Northwest Passage by *Gjøa* 1903–1905.

The French explorer Jacques Cartier entered the Gulf of St Lawrence and the St Lawrence river in 1535, which proved to be an entrance for France to the interior of North America, but found no passage to the Pacific. The English search

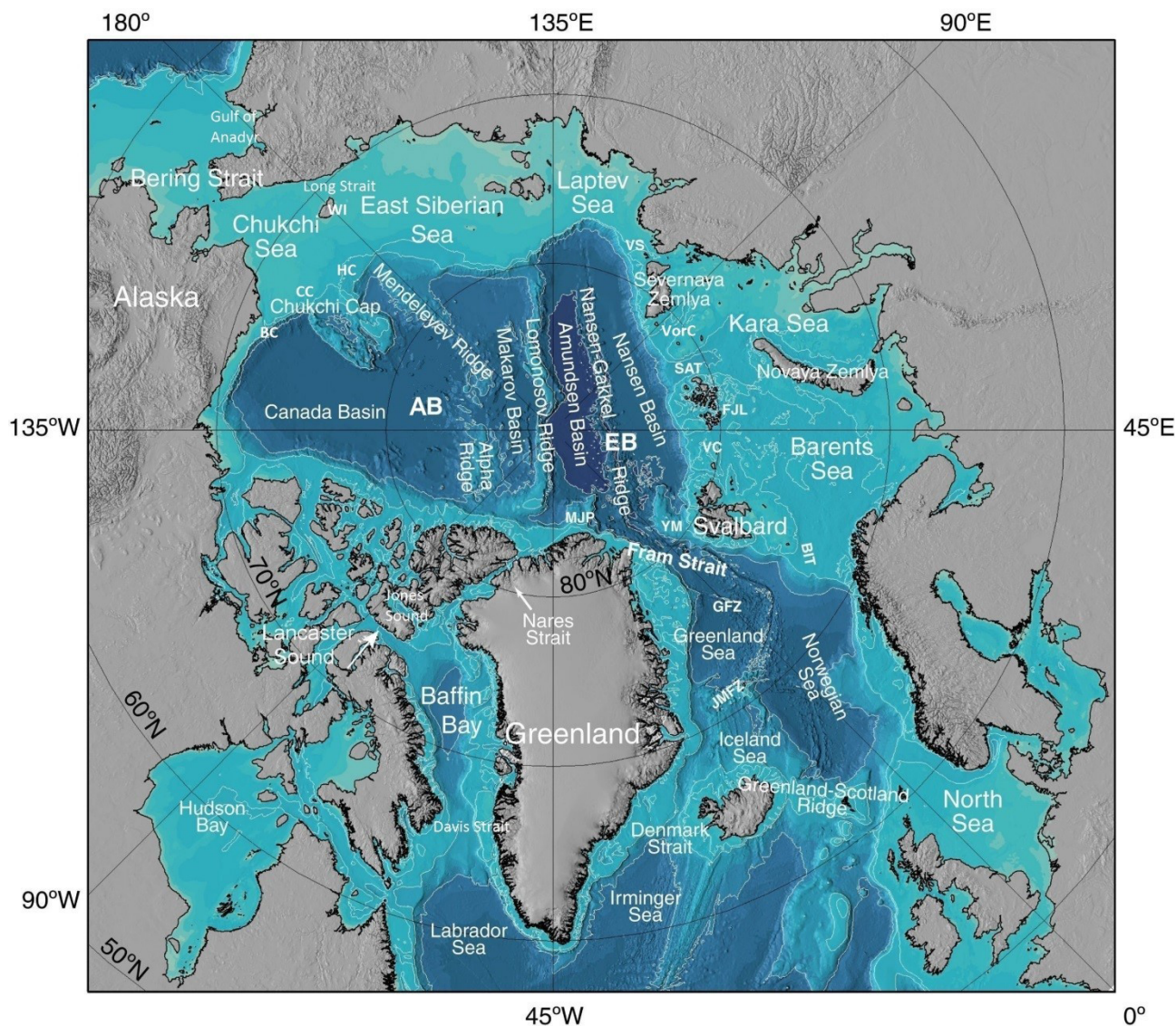


Figure 1. Map of the Arctic Mediterranean Sea showing geographical and bathymetric features. The bathymetry is from the IBCAO updated data base and the projection is Lambert Equal Area. The 200, 500, 2000 and 4000 m isobaths are shown. AB: Amerasian Basin, BIT: Bear Island Trough (Barents Sea opening), 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, YM: Yermak Plateau, VC: Victoria Channel, VorC: Voronin Canyon, VS: Vilkitskij Strait, WI: Wrangel Island (Adapted from Rudels, 2019, Fig. 1).

eventually turned to higher latitudes. In 1576 Martin Frobisher sailed west of Greenland and entered a passage that he thought might lead to the Pacific, but it was just a bay, later named Frobisher Bay. John Davis made two voyages west of Greenland and north of Frobisher Bay and on his return voyage he noticed a strait with strong currents south of Frobisher Bay that could indicate a passage west. In 1610 Henry Hudson was given the task to explore this passage, but after sailing through the strait, later named after him, and entering a large open water area autumn was approaching. The ship became trapped in the ice and the expedition had to overwinter. In spring when the ship came loose of the ice there was a

mutiny and Hudson, his son and seven members of the crew were put in a lifeboat to face a certain death.

As the ship returned to London the crew presented Hudson's sailing notes and was pardoned. Further expeditions showed, however, that Hudson had sailed into a bay, appropriately now named Hudson Bay, not into a passage to the Pacific. In 1616 Robert Bylot, with William Baffin as pilot, instead sailed as far north as they could west of Greenland and discovered three possible passages, Smith Sound, Jones Sound and Lancaster Sound but they judged that these were not leading to the Pacific. Baffin's sailing notes were, un-

fortunately, never published and the expedition was all but forgotten.

The English efforts to find the Northwest Passage eventually led to the establishing of one early successful commercial venture in the new world, the Hudson Bay company, and the search for the Northeast Passage resulted in contacts with Russian traders at the river Dvina and to the formation of the Muscovy company. This put an end to the English efforts to navigate the Northeast passage but the search continued, now mainly sponsored by Dutch merchants. On three voyages in 1594, 1595 and 1596 with Willem Barentsz as pilot the Dutch reached Novaya Zemlya but could not penetrate farther east. The 1596 expedition took a more northerly route and sighted Bear Island and Spitsbergen before it turned east and reached the northern cape of Novaya Zemlya. Here the ship was trapped in the ice and the expedition members overwintered on land. The following spring they tried to reach the Kola peninsula in a small boat, but Barentsz died already at the beginning of the sailing.

The discovery of the northern seas with their abundance of life, whales and walruses, was found more rewarding than failed searches for an unknown Northeast Passage and an intensive whaling industry was established, mainly by the Dutch, on Spitsbergen. Holland had now become independent from Spain and Spain had given up its monopoly of the southern routes to Cathay. No pressing need for the Dutch to find the Northeast Passage then remained (Fig. 2).

Russian explorers, however, continued to chart the northern Eurasian coastline. The northernmost cape, Cape Chelyuskin was reached in the early 17th century and in 1648 Semen Dezhnev sailed from the mouth of the Kolyma river in the Arctic Ocean to the Gulf of Anadyr in the Pacific, thus proving that the Northeast passage did exist and that Eurasia was separated from North America, almost a century before Vitus Bering sighted the Alaskan shore.

In the 18th century the initiative to finance expeditions to the northern seas shifted from merchants and adventurers to governments, and the expeditions focused on discoveries and on extending the influence of the home country. In 1730 Russia began its Great Northern Expedition led by Vitus Bering, where different segments of the Northeast passage were charted separately. Vitus Bering had on his first expedition observed, and on his second expedition he reached Alaska, but then died on the return voyage. Russia eventually established a permanent presence in Alaska, and this Russian settlement reactivated the English efforts to find the mythical Strait of Anian (Fig. 2) and the Northwest Passage, partly in concern that the Russians might be the first to discover the passage.

Bering had not completed the charting of the strait between Asia and North America. This was done by James Cook on his last voyage in 1779, and a later expedition, 1791 to 1795, commanded by George Vancouver, finally proved that the Strait of Anian did not exist. Any ship sailing into

the Arctic Ocean from the North Atlantic to the North Pacific would have to exit through Bering Strait.

About the same time the possibility to sail directly across the Arctic Ocean to Bering Strait was considered. One early, but failed, attempt to cross the Arctic Ocean had been made by Henry Hudson in 1608. In spite of the fact that all previous expeditions had been stopped by ice, the existence of an ice free inner part of the Arctic Ocean was seriously discussed. Robert Thorne, an English merchant, had already in 1527 suggested that the continuous daylight during summer should store enough heat in the water column to prevent ice formation in winter. Another idea, favored by the Dutch map-maker Peter Plancius, was that an ocean in perpetual motion would not freeze. The Russian scientist Michail Lomonosov suggested that ice can only form in connection with land and river runoff and the interior of the Arctic Ocean might be free of ice. In 1772 the Swiss geographer and librarian Samuel Engel noticed that sea ice contains little salt and proposed that sea water could not freeze. These ideas led to a few failed attempts to sail to the Pacific via the North Pole.

3.2 The deep circulation

In the age of sail the separation of the direct driving by the wind from those of the ocean currents was not easy to make. James Rennel studied the ship logbooks to determine the drift caused only by the ocean currents. The results of his analyses were published in 1832, 2 years after his death, as *Investigations of the currents of the Atlantic Ocean and of those which prevail between the Indian Ocean and the Atlantic* (Rennel, 1832) and supported the prevailing view that ocean currents are mainly wind driven. However, Mathew Maury made a similar compilation of ocean current from logbooks, which was published in 1855 as *“Geography of the Sea”* and in an edited edition eight years later (Maury, 1863). Maury, by contrast, argued that differences in density between water masses, due to temperature and salinity variations, might generate stronger currents than the wind.

In 1751 captain Henry Ellis on the vessel *Earl of Halifax* made temperature measurements in the deep tropical Atlantic by collecting water from different depths in a bucket equipped with a thermometer. He found that the temperature decreased to about 9 °C at 1200 m but then stayed constant, as the measurements continued down to 1600 m. These observations, although reported in the *Philosophical Transactions of the Royal Society*, did not receive much attention until Benjamin Thompson (Count Rumford) almost 50 years later proposed that the cold deep water was advected from higher latitudes, where the winters were cold enough to create water of the observed temperature. Thompson (1798) also pointed out that ocean water does not attain its maximum density at 4 °C, as in freshwater, but instead its density increases with decreasing temperature until freezing.

This was demonstrated by Emil von Lenz, who made deep temperature measurements using thermometers protected in

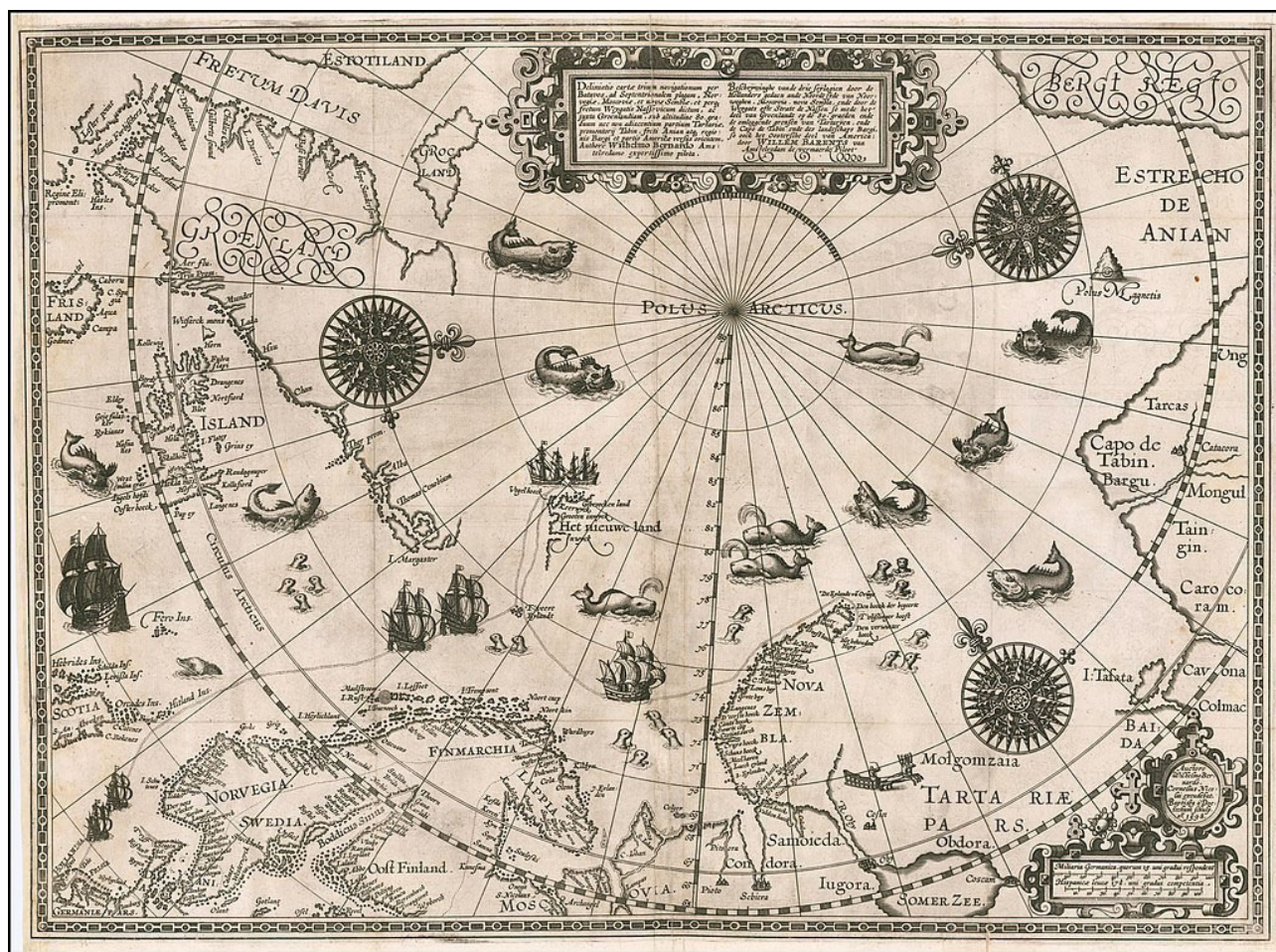


Figure 2. A map of the Arctic Ocean by Willem Barents drawn around 1598 showing the then known features of surrounding lands and islands. The ships north of Svalbard probably indicate open water, the Whalers' Bay (from Ericsson et al., 1992).

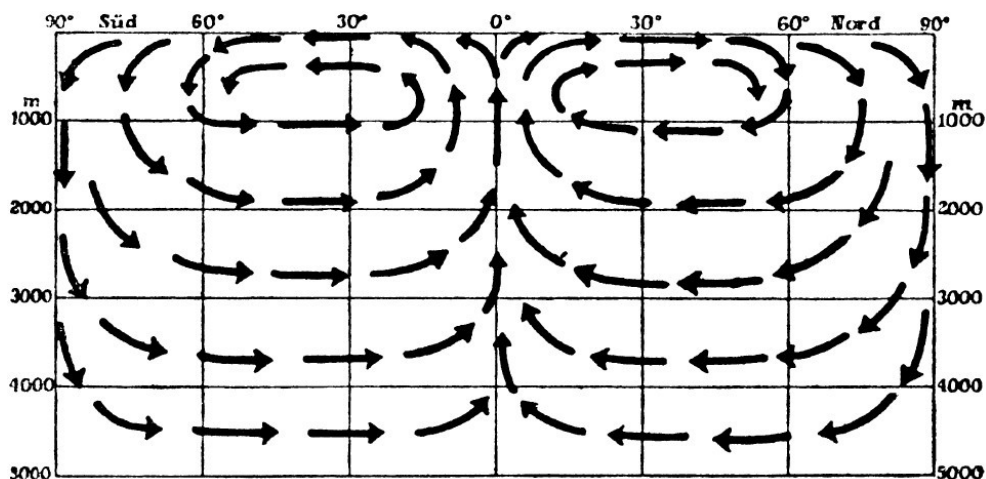


Figure 3. The symmetrical thermally driven meridional overturning circulation suggested by Emil von Lenz 1845 (from Merz and Wüst, 1922).

self-closing buckets on Otto von Kotzebue's second circumnavigation in 1823–1826, and observed temperatures well below 4 °C. Von Lenz returned to the study of the ocean deep circulation later in his life and proposed a thermally driven deep circulation symmetric about the equator, the water sinking at high latitudes and rising towards the surface at the equator (von Lenz, 1845) (Fig. 3).

In 1868 and 1869 William B. Carpenter and Charles Wyville Thomson sailed to the area between Scotland and the Faroe Islands to prove, or disprove, Edvard Forbes' theory of an ocean devoid of life below 800 m depth. They did prove that living creatures existed at the ocean bottom regardless how deep they sampled. However, they also found deep water with temperatures between 0 and 0.5 °C northeast of a line between the Shetland and the Faroe Islands, while the deep temperatures west of the line were between 4.5 and 8.5 °C. These observations suggested that waters cooled at higher latitudes were moving south along the bottom into the North Atlantic, and Carpenter (1871) proposed an extensive meridional overturning circulation, where cooling at the surface forces the surface water to sink at high latitudes to be replaced by warmer, lighter water advected in the upper layer from the equatorial region, while the dense deep water moves towards the equator, much like the circulation proposed by von Lenz (1845) and in agreement with the ideas of Benjamin Thompson.

Carpenter was aware of warm and strong surface currents such as the Gulf Stream but he considered these to be of minor importance and assumed that the effects of the Gulf Stream ended in mid-Atlantic and did not influence the climate in Europe. The overwhelming part of the heat carried northward by the ocean was instead due to the proposed massive, thermally driven overturning circulation. Carpenter's grand generalization of the meridional ocean circulation was initially not generally accepted and the main opposition came from James Croll.

Croll did not believe that heating of the surface water in the equatorial areas and cooling at high latitudes could create a sea level slope steep enough to drive the water to higher latitudes. He argued that the ocean currents carrying heat to the high northern and southern latitudes were wind driven, and in the North Atlantic this circulation was dominated by the Gulf Stream (see e.g., Croll, 1875). This debate lasted over five years but eventually abated. Croll returned to his theory of the ice ages (Croll, 1875) and Carpenter became involved in the *Challenger* expedition 1872–1876. Carpenter's thermally driven overturning circulation appeared in the end to have gained the strongest support. However, the question of how ocean currents, at the surface and in the deep, are driven was far from solved, and its complexity hardly appreciated by either Carpenter or Croll (Mills, 2009).

3.3 The Nordic Seas

At the time of the *Challenger* expedition the Norwegian biologist G. O. Sars and the meteorologist Henrik Mohn proposed to the Norwegian government to send an expedition, akin to the *Challenger* expedition, to the waters west of Norway, extending to Iceland and Greenland and northward to Svalbard and northeastward into the Barents Sea. The proposal was accepted and resulted in the Norwegian North Atlantic expedition 1876–1878 with S/S *Vøringen* to the “North Ocean”, as Mohn would call it in his comprehensive final report on the cruises (Mohn, 1887). The expedition thus studied the area, where the formation of the cold deep water observed by Carpenter and Wyville Thomson was expected to take place.

Mohn had worked with the mathematician Cato Guldberg on the thermal theory of cyclones and had formulated the equations for the baric wind, where the pressure gradient is balanced by the Coriolis force and the wind blows parallel to the isobars. Mohn noticed that the circulation in the Nordic Seas resembled that of a cyclone and he tried to determine the density distribution in depth from the temperature and salinity measurements made on the expeditions. Assuming a level of no motion around 300 fathoms (1 fathom = 1.85 m) he found that the sea surface was concave, being lowest at the center and standing higher along the Norwegian coast, where the warm Atlantic water moved north, and along the Greenland coast where cold but less saline water flowed south. This corresponds to a cyclonic circulation around the Nordic Seas where pressure is balanced by the Coriolis force. Mohn also included the effects of the mean winds. Using air pressure observations from existing meteorological stations, he computed the geostrophic wind and then related the current speed empirically to the wind.

Mohn assumed that wind drives the surface water in the same direction, but as the water is moving on a rotating earth the current becomes deflected to the right (Northern Hemisphere), leading to a change in sea level slope that eventually balances the Coriolis force and again makes the current parallel to the wind. The wind effect was added to the computed sea level distribution and found to be larger than and reinforcing the sea level slope due to the density distribution (Mohn, 1887).

The salinity observations made on the expeditions were, however, not of the standard required for the computations Mohn attempted, but the cyclonic circulation present in the Nordic Seas was reproduced. Furthermore, temperature observations indicated that below 700 m the Nordic Seas were filled with water with temperatures below zero, and in Norwegian Sea southeast of Jan Mayen an almost isothermal water column with temperature of −1.2 °C was observed. The salinity was comparatively high, and Mohn suggested that here a cooling and sinking of Atlantic water down to the bottom took place (Mohn, 1880). The Norwegian Sea would

then be one area, where the water that filled the deeper layers of the oceans was produced.

Mohn's theoretical study was a first attempt to mathematically describe the oceanic circulation but it was not immediately followed up. When it was reexamined twenty years later, it was mainly its weaknesses that were noticed and its merits were not fully appreciated (Werenskiold, 1937). The inaccurate salinity determinations led to instabilities and to unrealistic circulation patterns and Mohn's description of the wind driven currents was criticized. His suggestion of convection and ventilation of the deeper layers by cooling of the Atlantic water were also almost forgotten, when Fridtjof Nansen (1906) described the same phenomenon.

3.4 Warm Currents and ice cover

Croll had emphasized the importance of the heat carried by the Gulf Stream for the European climate, but warm currents that enter the Arctic Ocean might also melt ice and lead to an open Arctic Ocean beyond the region influenced by continental runoff. One strong advocate for an ice-free inner Arctic Ocean was the geographer August Petermann, who examined the path of and the heat carried by the Gulf Stream and traced Gulf Stream water along the Siberian shelf as far as Bering Strait (Petermann, 1865). He also suggested that the best route to enter the inner part of the Arctic Ocean would be over the Barents Sea east of Svalbard.

Petermann was perhaps the most ardent believer in an open Arctic Ocean and of the importance of warm ocean currents, but he was not the only one. Captain Silas Bent (1872) argued that the pathways into the Arctic Ocean were in the wake of the warm currents, and Van Campen (1876) tried to convince the Dutch authorities to revive the Dutch Arctic explorations, which had been so intense in the 17th century, by sending expeditions following different warm currents into the Arctic Ocean.

However, there were also contrary views. Maury (1863) argued that since the southward flowing cold water was found at the surface and had low salinity, it was instead warm and saline water from the south that rose from the deep and brought heat to the surface, which melted the ice and kept the inner parts of the Arctic Ocean open. The low salinity mixture of sea ice and cold water then flowed south in the East Greenland Current and the Labrador Current. This would then lead to a circulation akin to that suggested by von Waitz (1755) (see below).

Efforts to reach the open Arctic Ocean were made at all passages where warm surface water was observed to flow north. Elisha Kane (1853–1855), Charles Hall (1866–1871) and George Nares (1874–1875) tried to enter the Arctic Ocean west of Greenland, Julius Payer and Karl Weyprecht sailed 1872 with the ship *Admiral Tegetthoff* towards the Arctic Ocean east of Svalbard, and George Washington de Long (1879–1881) entered through Bering Strait with *Jeanette* in 1879. All these efforts failed and on several expeditions the

ships were trapped in and eventually crushed by the ice and the expedition members had to man the life boats. Some made it back, some perished. The *Tegetthoff* expedition discovered Franz Josef Land but then the ship was caught in the ice and crushed. The expedition members made it back in the life boats.

The *Jeanette* was trapped by the ice north of Bering Strait and drifted with the ice for 21 months in no obvious direction. After almost 2 years *Jeanette* broke up and vanished beneath the ice. The expedition members were prepared and had put provisions, sledges and boats on the ice, and they started the journey south towards Siberia. They reached open water in early August and took to the boats, but a storm separated the three boats. One boat disappeared, one boat reached the Lena river and the men were saved. De Long's boat made it to the Siberian coast, but they did not meet anyone to help them finding food and all died by starvation. Three years later, in 1884, objects from the *Jeanette* expedition were found by inuits on an ice floe close to Julianehåb on the southwestern coast of Greenland, indicating that they had been carried by the ice from the wreckage across the Arctic Ocean to the Nordic Seas and along the east coast of Greenland down to and around Cape Farewell.

3.5 The drift of *Fram*

Henrik Mohn examined the drift of the *Jeanette* objects and proposed that there was a continuous stream from the coastal region of eastern Siberia across the Arctic Ocean. This idea was taken up by Fridtjof Nansen, who suggested that one possible way to enter the interior of Arctic Ocean would be to work with the ice and the currents by letting a ship be carried by the ice drift over the Arctic Ocean. The discovery of objects from *Jeanette* put a time scale on the ice movements across the Arctic Ocean and on the duration of such a drift.

Nansen's plan split the scientific community. Some scientists were in favor but many against, claiming it to be impossible. However, Nansen had the support of the Norwegian government and eventually the idea gained the acceptance by a large fraction of the scientists. The crucial part was the ship, *Fram*, which was built by the Norwegian master shipwright Colin Archer specifically designed to withstand the ice pressure. *Fram* was rigged as a three masted schooner with a support steam engine. The shape of the hull was smooth to avoid the ship being caught by the ice. Instead, it would be forced upward as a cork onto the ice. The rudder could be drawn up into the hull and the propeller was protected. The heating and ventilation were designed to avoid that moisture collected on the inner walls of the ship and then froze to ice in the cold winters. *Fram* was also a research vessel with laboratories to work on the observations and the samples collected during the drift. *Fram* was in fact the first drifting scientific ice station.

Fram sailed from Oslo in June 1893 and in September she had reached the New Siberian Islands, where she turned

northward into the ice and was tied to an ice floe. The drift, lasting until summer 1896, had begun. The first winter showed that *Fram* could withstand the ice pressure. Meteorological and hydrographic observations were made regularly and depth measurements indicated water depths close to 4000 m. The tacit assumption that the Arctic Ocean was shallow was wrong, however, and the possibility of finding large land masses in the Arctic Ocean was ruled out.

The *Fram* expedition proved that the Arctic Ocean is an ice covered ocean, but also that it was possible to enter the ice cover by moving with the ice, if the vessel was sturdy and well designed. The existence of an ice drift, and possibly also a current, directed from Siberia towards Greenland and Svalbard was demonstrated. The expedition also showed that the Arctic Ocean in its central part is a deep ocean. The warm Atlantic water, by many assumed to enter the Arctic Ocean and with its heat keep the Arctic Ocean open, was also observed, but below 150 to 200 m depth and separated from the ice cover by a cold, low salinity upper layer. What had not been considered before was the magnitude of the freshwater input to the Arctic Ocean, and the fact that sea ice melting on warm water creates a protective cold and low salinity melt water layer at the surface that inhibits further melting (Nansen, 1912, see also below).

The knowledge gained by the drift of *Fram* did not, however, remove the sea ice and the obstacle it was for further studies of the Arctic Ocean, and in the following forty years oceanographic studies were mainly confined to the open areas and the marginal ice zone during the summer season. Plans for a submarine expedition to the North Pole with the discarded submarine *Nautilus* were made in 1931, but failure with the diving rudders prevented the submarine to dive. The original plans then had to be abandoned and the observations were confined to the northern part of Fram Strait and the area north of Svalbard (Sverdrup and Soule, 1933). In the late 1930s the Soviet ice breaker *G. Sedov* was trapped in the ice and involuntarily repeated the drift of *Fram*.

A major step forward in the research of the interior of the Arctic Ocean occurred in 1937, when the Soviet Union established the first drifting ice station, NP-1, at the North pole. The station drifted rapidly south and the team was picked up by Soviet ice breakers in February 1938 in the Greenland Sea. After the Second World War new ice stations were launched, mainly by the Soviet Union, but also by the United States and Canada. The Soviet Union also organized airborne expeditions each spring covering the entire Arctic Ocean, where water sampling, ice thickness measurements and meteorological observations were made. The observations from the ice stations and from the airborne expeditions provided the knowledge of the ice drift, the ocean circulation and the water mass distribution in the Arctic Ocean as it existed around 1980, before the beginning of the ice breaker and the submarine expeditions and the revolution in remote sensing and in mooring techniques that have taken place the last 40 years (see e.g., Rudels, 2015, 2021).

The presence of an atmospheric high pressure cell over the Beaufort Sea that generates an anticyclonic circulation of the ice, and the existence of the Transpolar Drift that transports ice from Siberia towards Greenland became better known. Observations of the density distribution in the water column also showed that the wind field creates an anticyclonic geostrophic circulation in the upper layers centered at the Beaufort Sea. The movements of the deeper Atlantic layer were determined from the observed changes in TS properties of the Atlantic core, which showed that the core temperature decreased around the Arctic basin, being highest north of Svalbard and lowest north of Greenland, implying a cyclonic circulation of the Atlantic water around the Arctic Ocean (Timofeyev, 1960; Coachman and Barnes, 1961, 1962, 1963; Coachman and Aagaard, 1974).

3.6 ICES and technical and theoretical advances

The drift of *Fram* was part of the tremendously active period in oceanography starting with the *Challenger* expedition 1874–1876. Between 1899 and 1902 the International Committee for the Exploration of the Sea (ICES) was established with its secretariat in Copenhagen and the ICES laboratory in Kristiania (Oslo) with Nansen as director. One of the first task of ICES, actually the reason for its founding, was to coordinate a five years observation program, where the different member states should arrange oceanographic expeditions in all seasons in their adjacent waters, thus covering much of the northern North Atlantic, the Norwegian Sea, the Barents Sea, the North Sea and the Baltic.

As a response the Norwegian Fishery Board launched an extensive research program in the Norwegian Sea between 1900 and 1904 with the research vessel *Michael Sars*. The report on these expeditions, *The Norwegian Sea*, published by Bjørn Helland-Hansen and Fridtjof Nansen in 1909, described not only observations made within the program, but almost all available observations and measurements were included in the report. *The Norwegian Sea* discussed all the major water masses and currents and most of the major circulation features; the Atlantic water entering from the south and flowing northward in the Norwegian Atlantic Current, the Norwegian Coastal Current ultimately from the Baltic Sea and moving along the Norwegian Coast, the southward flowing East Greenland Current and its branches the Jan Mayen Current and the East Iceland Current, and the major cyclonic gyres in the different basins were all identified. Not only surface currents but also the distribution and motions of the waters in the deeper layers were deduced and described.

Observations made in the North Atlantic from RV *Michael Sars* in 1910 (Helland-Hansen, 1912) and from *Frithjof* 1910 (Nansen, 1913) showed that the North Atlantic drift brings three branches of warm water towards the eastern part of the Greenland–Scotland Ridge, and two branches cross the ridge. One branch joins the slope current west of Scotland and enters the Arctic Mediterranean through the Faroe–

Shetland Channel, where the most persistent, warmest and most saline inflow takes place along the continental slope west of Shetland. This inflow then partly becomes a continuation of the European Slope Current, which is observed at the continental slope west of Spain and may even be present also farther south (Nansen, 1913). The second branch crosses the ridge west of the Faroes and flows first east and then moves south into the Faroe–Shetland Channel, where it splits, one part turning northward into the Arctic Mediterranean and the other circulating around the Faroe Islands. The third branch turns west along the ridge and flows south of Iceland into the Irminger Sea, where the main part continues south along the Greenland slope as the Irminger Current and finally enters the Labrador Sea. A small fraction separates from the stream and flows northward into the Iceland Sea west of Iceland as the North Icelandic Irminger Current, described by Martin Knudsen (1899) based on observations from the Danish *In-golf* expeditions 1895 and 1896.

The main function of the ICES laboratory in Kristiania was to develop, test and evaluate instruments and methods, formulas and tables to assure that reliable and repeatable measurements could be made. A goal that was actually reached in the first decade of the 20th century. The use of water bottles equipped with protected and unprotected reversing thermometers, and the determination of chlorinity by Mohr–Knudsen titration and calibrated with standard sea water provided by ICES became routine, and the salinity and density were calculated from the Knudsen Hydrographic tables (1901). This routine lasted for more than fifty years.

At the turn of the century Vilhelm Bjerknes formulated the circulation theorem for a stratified fluid on a rotating earth, which became the basis for the dynamic method by which the geostrophic velocities, arising from the balance between the Coriolis force and the pressure gradient can be determined (Bjerknes, 1898a, b, 1901). How to use the circulation theorem to compute ocean currents was described by Sandström and Helland-Hansen (1903) and Helland-Hansen (1905). Helland-Hansen also introduced the TS diagram as a method for describing water masses and stratification in the ocean (Helland-Hansen, 1916). Another advance, based on Nansen's observations on *Fram* that the ice drifted to the right of the wind, was the formulation of the theory of wind driven ocean currents by Vagn Walfrid Ekman (1905).

3.7 Deep water formation and the overturning circulation

In 1901 Roald Amundsen made an extensive survey of the waters in the Greenland Sea, northern Norwegian Sea and the Barents Sea onboard *Gjøa*, later famed as the first vessel sailing the Northwest Passage. These measurements together with other available observations, especially from Russian cruises in the Barents Sea (Knipovich, 1905), were compiled and interpreted by Nansen in his monograph *Northern Waters* (Nansen, 1906). The observations in the central Green-

land Sea showed that the stability of the water column was weak and an almost homogenous dome of cold, -1.2°C , and saline, 34.92, water with only a thin layer of warmer and less saline water separating it from the atmosphere. In winter seasonal cooling might increase the density of the surface water sufficiently for it to sink and renew the water of the deep water dome.

The observations in the Barents Sea close to Novaya Zemlya, by contrast, showed highly saline and cold bottom water, which Nansen interpreted as a result of brine rejection during ice formation. Nansen thus identified two modes of convection; homogenization of the water column by cooling in the open ocean, and the creation of dense saline water by brine rejection on the shelves. Nansen (1912) examined the possible sources of the cold deep water in the North Atlantic and concluded that dense water formed in the Nordic Seas might contribute to the North Atlantic deep water but he expected the main source to be located south of Greenland in the Irminger Sea. This statement by Nansen may have led to the view that the main transformation of the Atlantic water in the Arctic Mediterranean was into less saline upper water, resulting in too weak exchanges between the North Atlantic and the Arctic Mediterranean (see, e.g., Sverdrup et al., 1942).

Sandström (1908) made laboratory experiment with saline water in a narrow tank to describe the thermally driven circulation in the North Atlantic (Fig. 4). He found that the circulation, if only driven by cooling at the surface, was sluggish, and when cooling was applied at both ends of the tank cold water filled the main depth of the tank and the circulation was confined to a thin upper layer. However, if a heat source was applied at one end and a cold source at the other end of the tank a strong overturning circulation was generated, when the warm source was placed deeper than the cold source. By contrast, if the heat source was above the cold source the water column became stagnant. Since in the ocean the heat and cold sources are both located at the surface, this suggests that a thermally driven circulation is likely to be weak.

The density of sea water is determined not only by temperature but also by salinity (and by pressure). This allows for a different circulation mode. In 1755 J. S. von Waitz, a German mining engineer, estimated that evaporation in the Mediterranean Sea leads to a lower sea level, which forces a surface inflow of Atlantic water through Gibraltar Strait (von Waitz, 1755). He noted that evaporation removes freshwater but leaves the salt in the water column, which creates a more saline and denser water column in the Mediterranean Sea. This reverses the pressure gradient in depth and gives rise to a dense deep outflow to the Atlantic.

Von Waitz was aware of the two-layer exchange through the Bosphorus studied by Luigi Marsigli (1681), which showed less saline water from the Black Sea flowing over a denser deep inflow from the Mediterranean Sea. Von Waitz extrapolated this and his own findings to the world ocean. Heating and evaporation at lower latitudes would increase

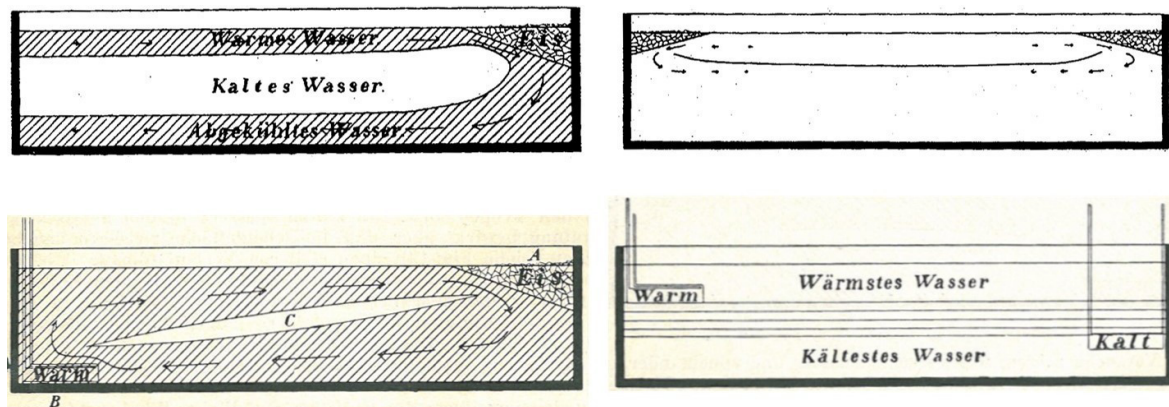


Figure 4. The Sandström experiment to simulate the circulation between lower and higher latitudes driven by cooling at high latitudes. The upper panel shows the effect by just cooling, at the Northern Hemisphere (left) and at both hemispheres (right). The circulation is sluggish. The lower panel shows the circulation with the cold source located above the heat source (left), leading to a vigorous circulation and the situation when the cold source is located below the heat source resulting in stagnant conditions (from Sandström, 1908).

the salinity and density of the surface water and force it to sink and be replaced at the surface of a flow of colder, less saline and less dense water from higher latitudes. This would create an overturning circulation in the opposite sense to that proposed by von Lenz and Carpenter.

Merz (1922) and Merz and Wüst (1922) reexamined temperature and salinity observations made in the South Atlantic on the *Challenger* expedition 1872–1876 and found that they did not agree with the symmetric circulation scheme proposed by von Lenz and that both the temperature and the salinity influence the circulation. Merz and Wüst (1922) emphasized that theirs were no new findings but had been noticed already in the original *Challenger* reports. The observed deep circulation does not consist of two loops symmetric about the equator but of several layers of sinking and advected water masses that extend from high latitudes across the equator into the other hemisphere. Cold, low salinity water around 1000 m depth moves from the South Atlantic towards the equator, while at 3000 to 4000 m depth a more saline and colder layer flows south from the North Atlantic, crossing the equator. The colder, slightly less saline bottom water originates in the south, in the Antarctica, and moves northward crossing the equator (Fig. 5). The deep circulation in the Atlantic thus forms a large meridional overturning circulation comprising several different layers that extend over the entire Atlantic connecting north and south and ultimately the North Atlantic to the rest of the world ocean.

4 The inflows and their forcing

The Arctic Mediterranean Sea is dominated by the interplay between the advection of heat, mass and freshwater from lower latitudes and the local, strongly seasonal, forcing. Most heat is carried by the atmosphere, which also provides the largest freshwater input to primarily the Arctic Ocean, mostly

as strongly seasonal runoff from the large catchment areas but also as direct local precipitation (Aagaard and Carmack, 1989; Carmack et al., 2015).

An oceanic inflow to the Arctic Ocean takes place through the shallow and narrow Bering Strait and in the *Oceans* H. U. Sverdrup estimated this inflow to 0.3 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) (Sverdrup et al., 1942). Russian work, however, indicated stronger transports and a significant seasonal variability, 1.2 Sv in summer and 0.4 Sv in winter (Maksimov, 1945; Fedorova and Yankina, 1964). The first direct current measurements made in and north of the strait suggested larger transports, 1.0 to 1.5 Sv, but no seasonal variability was detected (Coachman et al., 1975). The observational work in the strait has continued and later measurements vindicated the early Russian results. The transport varies seasonally with a larger northward transport in summer of 1.2 Sv and a smaller transport, 0.4 Sv, in winter (Coachman and Aagaard, 1981, 1988), and interannually (Itoh et al., 2026).

It has long been known that the northward transport through Bering Strait is due to the higher sea level in the North Pacific than in the Arctic Ocean when adjusted to a baroclinic level of no motion (Coachman and Aagaard, 1966), but Stigebrandt (1984) took a more global view and related the higher sea level in the North Pacific to the less dense Pacific water column, compared to that in the North Atlantic. A difference that ultimately arises from the atmospheric flux of water vapor from the Atlantic to the Pacific Ocean across the Isthmus of Panama (Weyl, 1968). However, also more local processes play a role. The weaker northward transports in winter have been explained by the presence of stronger, more northerly winds (Coachman and Aagaard, 1981, 1988). The inflow across the broad and shallow Bering and Chukchi shelves in summer is warmed, creating the Bering Strait summer water, while the cooling of the inflow in winter forms the Bering Sea winter water, which

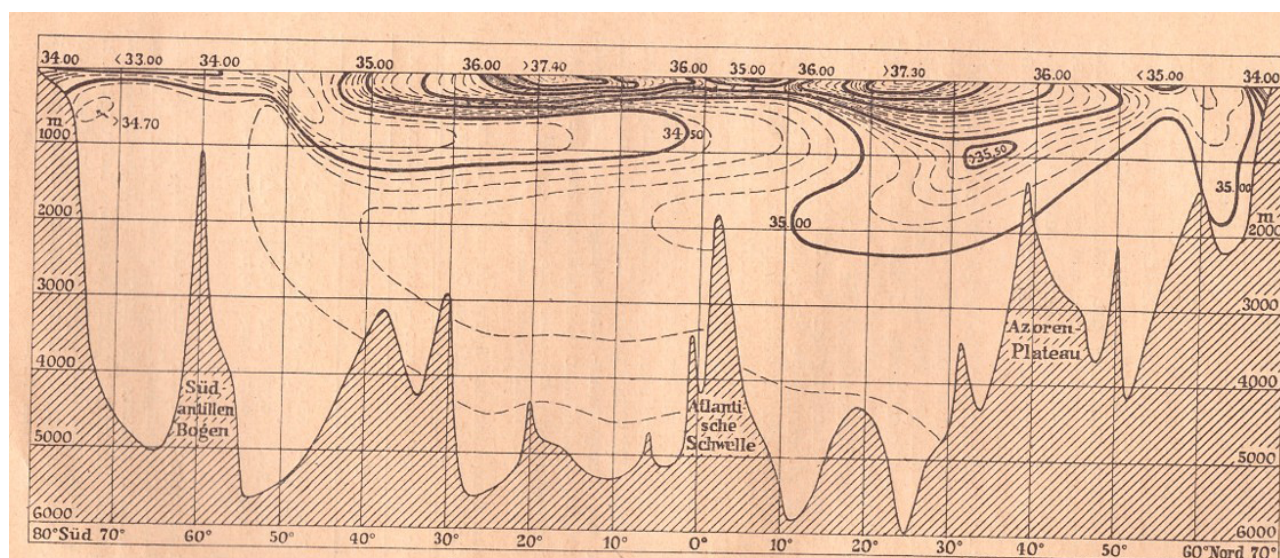


Figure 5. A longitudinal salinity section in the Atlantic showing the presence of the upper low salinity input from the Southern Ocean, the saline North Atlantic deep water and the slightly less saline bottom water from the Antarctic (from Merz and Wüst, 1922).

shows up as a temperature maximum and a deeper temperature minimum, respectively in the Canada Basin water column (Coachman and Barnes, 1961). The Pacific inflow is limited by the shallow sill depth (53 m) of the strait, and thus it does not carry oceanic heat from lower latitudes. Instead, the warmer water transported in summer is formed by local seasonal heating and is a part of the radiation balance of the Arctic Mediterranean. Because of its low salinity the Pacific inflow transports a substantial amount of freshwater to the Arctic Ocean, greatly adding to its stability.

By contrast, the connection between the Arctic Mediterranean and the North Atlantic over the Greenland–Scotland Ridge is deep and wide with a mean sill depth of about 500 m. The thickness of the Atlantic water layer entering the Norwegian Sea is 300–400 m and it transports oceanic heat from lower latitudes to the Arctic Mediterranean. The Atlantic water flows north as the Norwegian Atlantic Current in two branches. One follows the continental slope, while the second branch separates from the slope at the Vøring Plateau and continues as a baroclinic current above the Mohn–Knipovich Ridge towards Fram Strait.

As the Atlantic water at the slope reaches the Barents Sea opening it splits. One branch enters the Barents Sea but then again splits. The major part continues eastward toward Novaya Zemlya, while a smaller stream turns north in the Hopen Deep, where it again splits with one part flowing eastward north the Central Bank and the other returning as cold bottom water to the Norwegian Sea south of Bear Island. The Norwegian Coastal Current that flows between the Atlantic water and the coast, carrying low salinity water from the Baltic Sea and runoff from the Norwegian coast, also enters the Barents Sea and continues eastward close to the coast as the Murman Current.

The branch of Norwegian Atlantic Current that remains at the slope continues northward to Fram Strait, where it partly merges with the outer Mohn–Knipovich branch. Both branches then split. One stream enters the Arctic Ocean along the continental slope close to Svalbard, and a second stream flows around the Yermak Plateau to rejoin the inner branch farther east. A smaller eastward flow crosses the Nansen Ridge that connects the Yermak Plateau to the Svalbard slope into the Sofia Deep (Gascard et al., 1995). A substantial fraction of the Atlantic water, however, recirculates westward and does not enter the Arctic Ocean. Instead, it flows south along the East Greenland shelf and slope, as noticed early by Carl Ryder (Ryder, 1892). The cooling of the Atlantic water in the Norwegian Sea increases its density sufficiently for it to become overflow water that could contribute to the deep return flow of the North Atlantic overturning circulation.

In contrast to Bering Strait the mechanisms driving the Atlantic water into the Nordic Seas are not well known. The warm Atlantic water derives from the Gulf Stream and from the Subtropical gyre. As the warm water in the North Atlantic Drift approaches the British Isles it splits. The upper layer moves mainly south and remains in the subtropical gyre, while the deeper layers, below 300 m, rise to the surface and flow north into the Subpolar gyre (Brambilla and Talley, 2006; Burkholder and Lozier, 2011). The functioning of this valve and how it might influence the properties of the northward flowing Atlantic water is not adequately known but likely depends on a combination of the characteristics of the North Atlantic Drift, the circulation in the Subpolar gyre and on the properties of the waters exiting the Arctic Mediterranean.

The more fundamental question is if the inflow across the Greenland–Scotland Ridge is driven by winds and large

weather systems like the North Atlantic oscillation, or if the cooling of the Atlantic water and the convection taking place in the Nordic Seas draw the warm Atlantic water across the ridge in the manner envisaged by Carpenter? This will have a large impact on how the Arctic Mediterranean responds to changes in climate and we return to this question later.

5 The ice cover and ocean ice atmosphere interactions

The most outstanding feature of sea ice is that it floats. It is not unique but rare that a substance, when solidified by cooling, becomes less dense than its liquid phase. In freshwater this is preceded by a density maximum at 4 °C, above the freezing point, which means that further cooling creates its own stable stratification, limiting the amount of water that must reach freezing temperature before ice is formed. Sea water with salinity $S > 24.7$, by contrast, is densest at its freezing point and the water column has to be convectively cooled to freezing temperature down to the bottom, or down to a density gradient caused by the presence of a deeper more saline water mass, before sea ice is formed. As the water column is at the freezing point no sensible heat is supplied by the underlying water and the only heat source for the atmosphere is the release of latent heat of freezing, and a balance between the heat lost at the ice surface and the heat conducted through the ice is created. A balance that gradually shifts as the thickness of the ice cover increases and the ice surface temperature decreases, which reduces the ice growth at the ice-sea interface.

The solid ice cover allows snow to accumulate during winter, which isolates and further reduces the heat loss to the atmosphere and to space. A white snow cover implies a high albedo and when the sun rises in spring most of the shortwave radiation is reflected back to space. This is perhaps the most often mentioned Arctic feedback mechanism. Warmer air, less snow, more absorbed shortwave radiation, warmer air, less snow, etc. However, another mechanism that may be even more important is increased advection of moist air from lower latitudes, which increases the downward longwave radiation and may start melting snow and ice before sunrise (Mortin et al., 2018).

During summer the incoming solar radiation increases the air and water temperatures and snow and sea ice start to melt, creating an upper melt water layer that, however, remains close to melting temperature as long as ice is present. If the ice cover is not compact, shortwave radiation penetrates into and through the melt water layer and creates a warmer layer below. This heat is isolated from the ice until autumn, when ice again starts to form and the salinity of the melt water layer increases. These two layers might then merge, which would temporarily slow down the seasonal ice formation. If not, the warmer layer becomes a permanent temperature maximum in the water column, the Near Surface Temperature Maximum (Jackson et al., 2010, 2011).

A stable seasonal cycle of ice formation and ice melt would eventually lead to an equilibrium thickness of the ice cover. The equilibrium is reached because the ice formation in winter depends more upon the ice thickness than the ice melt in summer. Thinner ice leads to rapid growth in winter and the ice cover becomes thicker. If the ice is initially thick, little ice growth takes place in winter, while a larger amount melts in summer and the ice thickness decreases (Maykut, 1986).

As sea ice forms, either by growing directly from the underside of the ice cover or created as frazil ice in the water column, salt and freshwater become separated into brine and ice. The sea ice becomes thicker and the water below saltier, and in the case of frazil ice formation the ice crystals rise to the surface and “droplets” of brine sink. The smaller brine droplets soon lose their identity as they sink and the salt diffuses into the surrounding water. This creates a thicker, slightly unstable layer that eventually must convect into the underlying ocean (Rudels, 1986a). However, the frazil ice that is collected as a “mushy” layer (Wettlaufer et al., 1997) at the surface traps a large fraction of the brine, and as the mushy layer solidifies the brine is eventually squeezed out of the ice and sinks. The brine parcels rejected out of the mushy layer are likely larger than the droplets formed together with the frazil ice and might sink as individual parcels to greater depth before the salt anomaly is diffused away.

Until now we have only looked at the case when the upper layer temperature is close to the freezing/melting temperature and not considered what happens, when sea ice encounters and melts on warmer water. Nansen (1912) studied this situation by laboratory experiments. A vessel with warm, saline water was cooled from above in three different manners. First a jar with sea ice was held above the water surface and the water was cooled by radiation. Then the jar was allowed to float on the water, cooling the water by conduction. Finally, the ice was added at the surface and coming in direct contact with the ice (Fig. 6). The temperature evolution at the bottom and at the center of the vessel was measured in all experiments. The difference between the three experiments is striking. In the case of just radiative cooling the temperature decrease of the water is small and slow, while the conductive cooling of the water leads to a rapid and large temperature change in the vessel, indicating that thermal convection is established. The experiment where the ice is melting directly at the sea surface is the most revealing. A short rapid cooling takes place but after that the temperature remains essentially constant. The warm water melts the ice, which leads to the formation of a cold, less saline and less dense surface layer that isolates the ice from the warm water below. Sea ice evidently protects itself from heat input from below by creating a stratification that limits the heat transfer.

Nansen’s experiments were idealized. No turbulence was present in the vessel and the experiments were performed at room temperature. In the ocean wind will generate turbulence, either directly or by driving the ice, which entrains

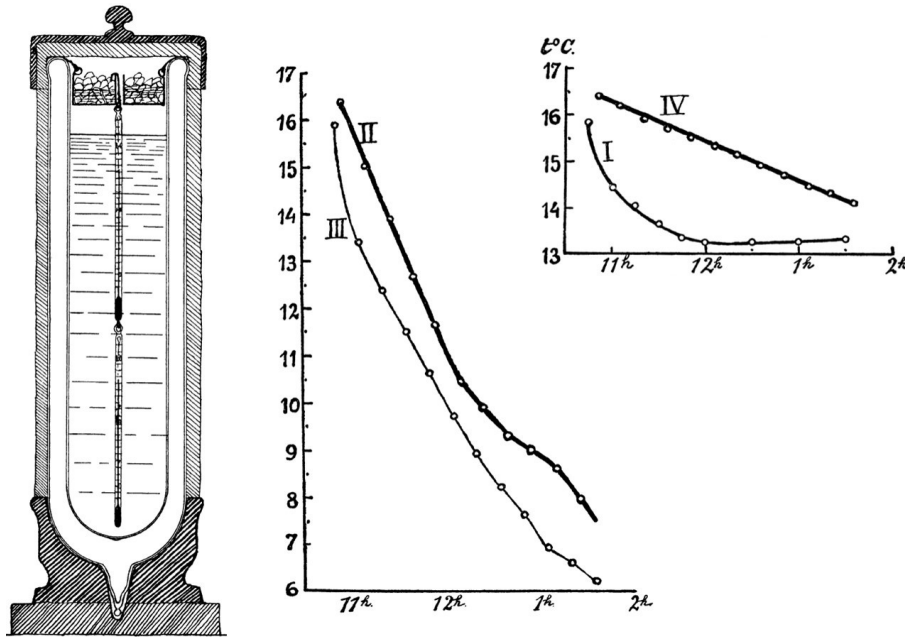


Figure 6. Nansen's experiments with cooling of sea water. Experiment I is when the ice is placed directly on the water, which leads to the formation of a low salinity melt water layer and no or little convection in and cooling of the water column. Experiments II and III show the results when the ice is put in a jar, which is floating at the water surface. The cooling at the surface leads to convection and cooling of the water column. In experiment IV the jar with ice is held above the water surface. The water is cooled slowly by radiation and the convection is weak (adapted from Nansen, 1912).

water across the interface into the upper layer. Furthermore, in summer the main sea ice melting occurs at the surface and the melt water layer prevents heat from below to reach the ice. In winter, however, the air temperature is low and the summer melt water has again become ice. As the warm water gets in contact with and melts the ice, an upper, cooler and less saline and less dense layer develops and deepens as warm water is entrained from below.

Assuming that the deepening of the upper layer can be described by a Kraus-Turner energy balance model (Kraus and Turner, 1967; Niiler and Kraus, 1977), where the turbulent energy added at the surface is used not only to entrain denser water from below but also to mix melt water from above into the mixed layer the entrainment velocity w_e becomes:

$$w_e = \frac{2n_0 u_*^3}{g[\beta(S_2 - S_1) - \alpha(T_2 - T_1)]} - \frac{\epsilon B}{g[\beta(S_2 - S_1) - \alpha(T_2 - T_1)]} \quad (1)$$

Here n_0 is a constant usually put to 1.25, u_* is the friction velocity, T_1 , T_2 , S_1 and S_2 are the temperatures and salinities of the upper and lower layer, respectively, α is the coefficient of heat expansion and β the coefficient of salt contraction. The sign of α is explicit in the expression. g is the acceleration of gravity, B is the buoyancy input and ϵ is a parameter put to 1 if the buoyancy input is positive and 0.05 if the buoyancy is negative to take the weak mixing rate of convection

into account (Stigebrandt, 1981). If the density is taken only to depend upon salinity, the second term becomes equal to w_e and the input of fresh (melt) water at the surface thus reduces the entrainment by a factor of 2 (Rudels, 1989).

The entrained heat not only melts ice, in winter the heat loss to the atmosphere also has to be supplied by the entrained water. The distribution of the entrained heat to sea ice melting and to the atmosphere is crucial for the evolution of the upper layer, and one suggestion is that the fraction of heat going to ice melt is the one that leads to minimum ice melt (Rudels et al., 1999a). This fraction ϕ_0 is found to be:

$$\phi_0 = \frac{2\alpha L}{c(\beta S_2 - \alpha \Delta T)} \approx \frac{2\alpha L}{c\beta S_2} \quad (2)$$

L is the latent heat of melting and c the heat capacity of sea water and ΔT the temperature difference between the Atlantic and the melt water layer. (Rudels et al., 1999a). α decreases with decreasing temperature and Fig. 7 shows the total heat loss Q' , the heat loss to ice melt $\phi_0 Q'$, and the heat loss to the atmosphere $(1 - \phi_0) Q'$ as functions of ϕ for different ΔT and the corresponding values of α .

As the created upper layer deepens its salinity and temperature decrease, while the stability increases with time until the freezing temperature is reached. The salinity S_1 in the upper layer is then given by:

$$S_1 \approx \frac{S_2}{\left(1 + \frac{2\alpha}{\beta S_2} (T_2 - T_f)\right)} \quad (3)$$

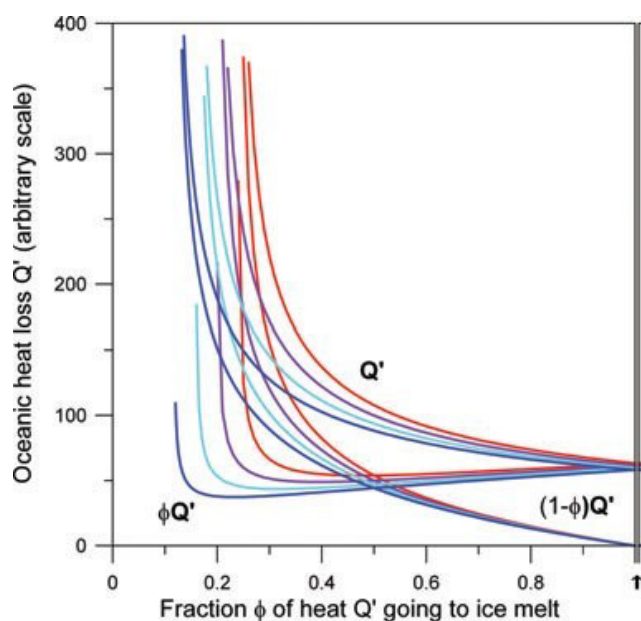


Figure 7. Curves showing the total heat loss Q' , the heat going to ice melt $\phi Q'$ and the heat going to the atmosphere $(1 - \phi)Q'$ as function of ϕ , the fraction going to ice melt. Red curves have $\Delta\alpha T = 8^\circ\text{C}$ and $\alpha = 0.781 \times 10^{-4}$, purple curves $\Delta\alpha T = 6^\circ\text{C}$ and $\alpha = 0.654 \times 10^{-4}$, cyan curves $\Delta\alpha T = 4^\circ\text{C}$ and $\alpha = 0.526 \times 10^{-4}$, blue curves $\Delta\alpha T = 2^\circ\text{C}$ and $\alpha = 0.390 \times 10^{-4}$. $\beta = 8 \times 10^{-4}$, $c = 4000 \text{ J kg}^{-1} \text{ K}^{-1}$, $L = 0.335 \times 10^6 \text{ J kg}^{-1}$. The fraction ϕ_0 giving minimum ice melt decreases with decreasing temperature and decreasing α (from Rudels, 2010).

T_2 is the temperature of the lower layer and T_f the freezing temperature of the upper layer. The stable salinity step $\beta(S_2 - S_1)$ is about twice the unstable temperature step $\alpha(T_2 - T_1)$ giving a stability ratio of 2 at the interface. The problem is the choice of α , which depends strongly on temperature and as the upper layer deepens and cools α decreases to a minimum at the freezing point. A smaller α leads to a deeper upper layer with higher salinity.

When the freezing temperature is reached, the upper layer has become so deep that the amount of entrained warm water is not enough to supply heat both to ice melting and to the atmosphere, and ice again starts to form. The entrained oceanic heat loss, however, is still expected to be distributed between ice melting and the atmosphere that the ice melting is a minimum. The stability at the interface now decreases due to ice formation and brine rejection and the upper layer eventually becomes dense enough to convect into the deep underlying water (Rudels et al., 1999a; Rudels, 2021). This situation will be discussed in more detail in Sect. 10 below. Here we might just add that the fraction ϕ_0 of minimum heat going to ice melt is twice the fraction leading to marginal stability. A fraction smaller than $\phi_0/2$ will not stabilize the water column and it overturns. This is also evident from the

stability ratio $R^p = \beta \Delta S / \alpha \Delta T = 2$ reached when the upper layer has attained freezing temperature.

6 The Atlantic inflows to the Arctic Ocean

The Atlantic water that enters the Arctic Ocean through Fram Strait soon encounters and starts to interact with sea ice. In winter the melting sea ice and entrained Atlantic water create a less saline melt water layer that is eventually cooled to freezing temperature (Rudels et al., 1996). As a result, an area of open water often forms north of Svalbard, the “Whalers’ Bay”. Whether heat entrained by mechanical mixing from the Atlantic layer below is sufficient to melt the ice and create a polynya, or if convection of the entire upper layer, driven by ice formation and brine rejection after it has reached freezing temperature, is also required, remains an open question. The Whalers’ Bay normally becomes ice covered later in winter, either by locally formed ice or by ice advected from the interior of the Nansen Basin.

The formation of the less saline upper layer implies that part of the Atlantic water that was cooled in the Norwegian Sea and transformed into high density water, which potentially could contribute to the overturning circulation, instead will leave the Arctic Mediterranean as a less dense upper outflow. The now stratified Atlantic water column continues as a boundary current eastward along the Eurasian continental slope and in the following summer sea ice melting by short-wave radiation creates a seasonal surface melt water layer.

The fate of the water that enters the Barents Sea is more varied and more difficult to follow. The Norwegian Coastal Current and the main Atlantic inflow branch flow eastward towards Novaya Zemlya and the heat advected by the Atlantic water is sufficient to keep the southern half of the Barents Sea free of ice the year around. The heat loss here is large and the Atlantic water becomes considerably colder than the Atlantic water entering the Arctic Ocean through Fram Strait. In the southern Barents Sea ice is only formed close to the coast and over the shallow areas west of Novaya Zemlya, which are mainly covered by water from the less saline coastal current and where the entire water column is cooled to freezing in winter.

The northern Barents Sea is different. It is shallow, mainly comprising the about 120 m deep Grand Bank. It connects to the Arctic Ocean between Svalbard and Franz Josef Land with the deepest trough being the 500 m deep Victoria Channel west of Franz Josef Land. The northern Barents Sea is strongly stratified and seasonally ice covered. The source(s) of the low salinity upper layer is(are) not obvious. Its characteristics suggest that it is not directly supplied from the Arctic Ocean, but it might enter from the Kara Sea in the Persey Current, which flows westward south of Franz Josef Land. Another possibility is that the upper layer is formed in the Barents Sea by sea ice melting on the warmer Atlantic and coastal waters that move north towards the pas-

sage to the Kara Sea between Franz Josef Land and Novaya Zemlya. The low salinity, cold upper water is eventually exported from the northern Barents Sea to the Norwegian Sea in the Bear Island Current. The more saline and denser deep and bottom waters in the northern Barents Sea are supplied by the branch of Atlantic water that mainly recirculates in the Hopen Deep, but partly penetrates northward onto the Grand Bank.

In the 2010s the ice cover in the northern Barents Sea was significantly reduced and the stability between the upper layer and the Atlantic water below weakened. This was suggested due to an increased inflow of warmer Atlantic water that reduced the stability and melted the sea ice from below (Lind and Ingvaldsen, 2012; Lind et al., 2018). The change would be part of a larger shift that is presently occurring in the Arctic Ocean, the so called “Atlantification” of the Arctic Ocean, which implies a reduction of the upper layer thickness and stability, increased shear and a stronger upward transport of heat from the Atlantic layer to the ice (Årthun et al., 2012; Polyakov et al., 2017; Årthun et al., 2025). This phenomenon was first reported from the eastern Nansen Basin north of the Laptev Sea, and the changes observed in the Barents Sea were taken as signs that the same pattern was expanding. The heat advected from the North Atlantic would in the near future dominate not only the Norwegian Sea but also the Nansen Basin and the Barents Sea and reduce the extension of the ice cover (Lind et al., 2018; Polyakov et al., 2020a, b). In the later part of the 2010s the reduction of the ice cover in the northern Barents Sea stopped and it started to recover, but it is still far from its extent in the 1980s.

Not only the ice cover but also the water mass characteristics in the Barents Sea have changed considerably during the last decades. In the 1980s the ice cover and the low salinity upper layer extended farther south and occasionally covered the Central Bank. In some years ice formation and brine rejection over the bank led to haline convection to the bottom and the formation of a cold, dense water column over the bank that eventually contributed to dense water formation in and dense water export from the Barents Sea (Quadfasel et al., 1992; Schauer et al., 2002). In years with strong inflow of warm Atlantic water the Central Bank, by contrast, became covered by Atlantic water and no ice and no dense water was formed over the bank. In the 2000s and 2010s the Atlantic water has been warmer, more saline and denser than in the 1980s. The convection over the Central bank has been thermal and the created waters have been denser than those formed by brine rejection in the 1980s.

The dense water formation in the Barents Sea has altered not only over the Central Bank but also farther east, above the Central Depression and over the shallow areas west of Novaya Zemlya. The Atlantic water cooled in winter is presently denser than in the 1980s and shows up as warmer bottom water below a less saline water mass with temperature close to freezing (Lien and Trofimov, 2013; Dmitrenko et al., 2015). The salinity indicates that the upper layer could

be created by sea ice melting on top of the denser Atlantic water (Eq. 3 above). This is in contrast to observations made in the early 1900s, when the water created by ice formation and brine rejection west of Novaya Zemlya was the densest water formed on the Barents Sea (Knipovich, 1905; Nansen, 1906). Dense water formation by brine rejection still occurs but now it appears to mainly take place north rather than west of Novaya Zemlya (Lien and Trofimov, 2013).

7 Features of the circulation in the Arctic Mediterranean Sea

7.1 Atlantic water in the Eurasian Basin

The two Atlantic inflow branches again meet in the St. Anna Trough in the western Kara Sea. The Fram Strait branch, flowing as a boundary current along the Eurasian continental slope, follows the bathymetry southward on the western side of the trough, while the Barents Sea branch enters the trough from southwest through the passage between Novaya Zemlya and Franz Josef Land. The properties of the two branches are now different. The Atlantic core of the Fram Strait branch has been shielded from the surface processes by the less saline upper layer created by ice melting north of Svalbard and has retained its warm and saline characteristics. The main part of the Barents Sea inflow branch, by contrast, has been strongly cooled by the atmosphere in the southern Barents Sea and when it encounters and melts sea ice in the northeastern Barents Sea the salinity and density of the melt water layer become higher than that created above the Fram Strait branch north of Svalbard (Eq. 3 above).

The part of the Fram Strait branch that flows close to Franz Josef Land enters the Barents Sea and circulates around Franz Josef Land to return to the Arctic Ocean in the Victoria Channel, rejoining the boundary current upstream. Most of the Fram Strait branch, however, interacts in the St. Anna Trough with water of similar density from the Barents Sea, forming a colder and less saline Atlantic core that remains on the slope and moves eastward between the Kara Sea shelf and the warmer Fram Strait branch further down the slope. The denser waters formed in the Barents Sea sink down the eastern flank and along the deepest part of the St. Anna Trough and enter the boundary current as cold, less saline intrusions beneath the warm core of the Fram Strait branch.

Two of the largest Siberian rivers, Ob and Yenisey, discharge into the Kara Sea, but most of their runoff does not appear to cross the Kara Sea shelf break. The major part of the river input instead flows through the Vilkitsky Strait to the Laptev Sea before it, together with runoff from the Lena, crosses the shelf break and enters the deep Eurasian Basin. This shelf outflow is less saline and less dense than the surface layer created north of Svalbard and increases the stability of the upper part of the water column. The Atlantic water becomes more isolated from the ice cover, the sea sur-

face, and the atmosphere and most of the observed property changes are due to interactions in the deep between the two inflow branches.

The upper, colder stream follows the shelf break until it reaches Severnaya Zemlya. Here the slope and shelf narrow and the stream is forced into the basin and starts to interact isopycnally with the warm core of the boundary current (Rudels et al., 1994; Schauer et al., 1997). Inversions and thermohaline intrusions are created between the colder, less saline shelf stream and the warmer, more saline slope current (Fig. 8). Intrusions are observed in all possible background stratifications; above the temperature maximum where the water column is unstable in temperature, between the temperature and the salinity maxima where the water column is stable in both properties, and below the salinity maximum where the water column is unstably stratified in salinity.

In the Amerasian Basin the Atlantic waters no longer provide a salinity maximum and thermohaline intrusions, the signature feature of the 2000's warming of the Atlantic layer (see below), only persist in the stable-stable part below the temperature maximum (Carmack et al., 1995; McLaughlin et al., 2009). Thermohaline intrusions also display a collective, nested structure in T/S space extending fully across the Arctic Ocean interior (Carmack et al., 1998; Walsh and Carmack, 2002, 2003).

Theoretically, instabilities and interleaving may form spontaneously, if one component is unstably stratified (Stern, 1967; Turner, 1973), but if the background stratification of both components are stable, finite external disturbances are needed to bring the waters across the front to generate the property inversions that drive the intrusions. At the slope such disturbances could be generated by trapped tidal or inertial waves (Toole and Georgi, 1981). However, the intrusions penetrate across the front into the opposite water masses only as long as an unstable distribution of the driving component remains. Once it is exhausted, the interleaving stops (Rudels and Hainbucher, 2020; Rudels, 2021). One question is then: How far from the slope into the basin do the intrusions influence the lateral mixing between the two streams?

Interleaving structures are observed at the Kara Sea slope but also in the interior of the Eurasian Basin at least as far as the Gakkel Ridge. However, a core of warm, saline Fram Strait branch water showing no or weak intrusions is located between these interleaving structures (Fig. 8). It thus appears that the intrusions formed at the slope do not reach directly to the center of the warm core. On the basin side of the warm core there is no obvious front that can generate intrusions. Furthermore, the only source of colder, less saline water in this density interval is the Barents Sea inflow branch.

One possible explanation for the presence of thermohaline intrusions in the central Nansen Basin is that the two branches, as they reach the Laptev Sea slope, turn and flow back towards Fram Strait in the Nansen Basin and over the Gakkel Ridge (Rudels et al., 1994). This implies that the intrusions do not spread as far as their presence suggests. In-

stead, the observed interleaving structures in the interior of the basin are rundown and fossil and are passively advected with the main circulation. In fact, they could be used as tracers for the circulation (Rudels et al., 1994).

The water column is not only characterized by interleaving and gyre circulation. There are also several eddies present, which appear to move independently of the gyres, carrying their original properties into different surroundings (Fig. 8, lower panel).

The interleaving between the two branches leads to cooling of the Fram Strait branch, but because mass is added the cooling does not necessarily imply heat loss to the overlying water and to the ice and atmosphere. The heat has already been lost elsewhere, mainly in the Barents Sea. Not only is the Atlantic core cooled, the salinity also decreases significantly (Fig. 9). To mechanically mix freshwater or low salinity water vertically into the deep is energy consuming, and the more obvious process is isopycnal mixing between the two branches. The low salinity water added from the Kara Sea west of Severnaya Zemlya enters, freshens and cools the Atlantic core in contrast to the low salinity water from the Laptev Sea, which enters as a surface layer and rather isolates the Atlantic water from the sea surface and the atmosphere.

This is revealed in Fig. 9, which shows the heat content (relative to the freezing point, -1.8°C) and the freshwater content (relative to salinity 34.9) on stations taken on three sections occupied by FS *Polarstern* in 2007 and 2011 from the Eurasian continental slope into the Eurasian Basin for six different water masses (Fig. 9). The total height of the columns indicates the total heat content above 1000 m, but only off the Laptev Sea does the column heights show the total freshwater content. On the Franz Josef Land and the Severnaya Zemlya sections the freshwater content in the Atlantic waters is negative and has to be subtracted from the water column heights. This shows the substantial volume of freshwater that is added to the deeper Atlantic layers between the Kara Sea and the Laptev Sea, confirming the importance of the isopycnal mixing between the branches.

To understand how present, and especially how possible future changes in inflow strength and characteristics of the Atlantic water might influence the condition in the eastern Nansen Basin, it is necessary to consider both inflow branches, how they are transformed individually and how they interact.

Interleaving has vertically alternating salt finger and diffusive interfaces and between the background temperature and salinity maxima the buoyancy transports through the salt finger and diffusive interfaces are roughly equal and the property steps diminishes with time but are not removed. However, if intrusions are created in the upper part of the water column, above the temperature maximum, the diffusion of buoyancy through the diffusive interfaces is larger than that through the salt finger interfaces due to the larger property steps created at the diffusive interfaces. This may lead to overturning of the salt finger interfaces, which transforms the

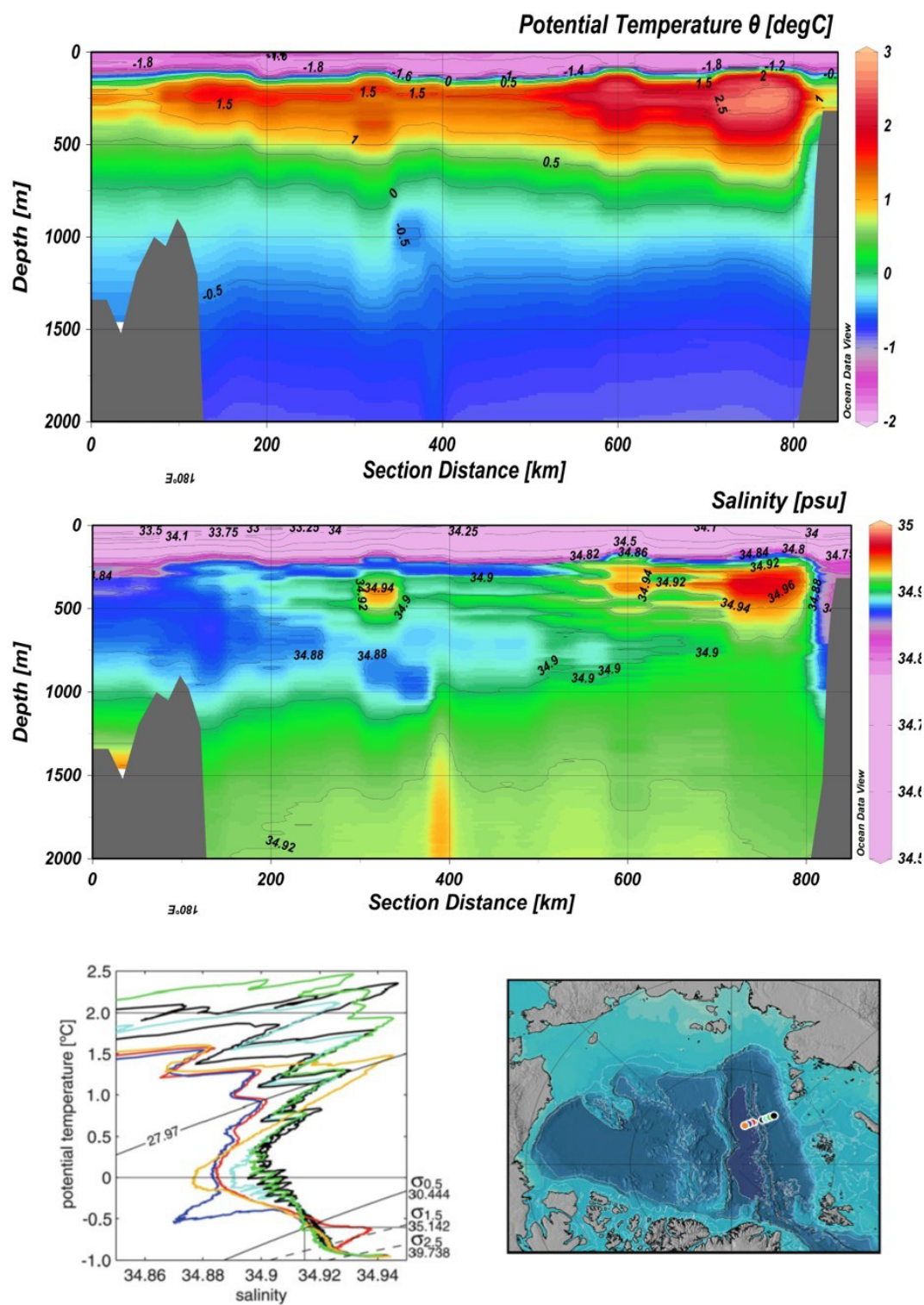


Figure 8. Potential temperature and salinity sections taken in 1996 by FS *Polarstern* from the eastern Kara Sea shelf across the Nansen and Amundsen basins and the Lomonosov Ridge into the Makarov Basin. The θ - σ curves are from the same section, showing interleaving as well as eddies in the different waters. The station positions are shown on the map (The lower panel is from Rudels et al., 2012).

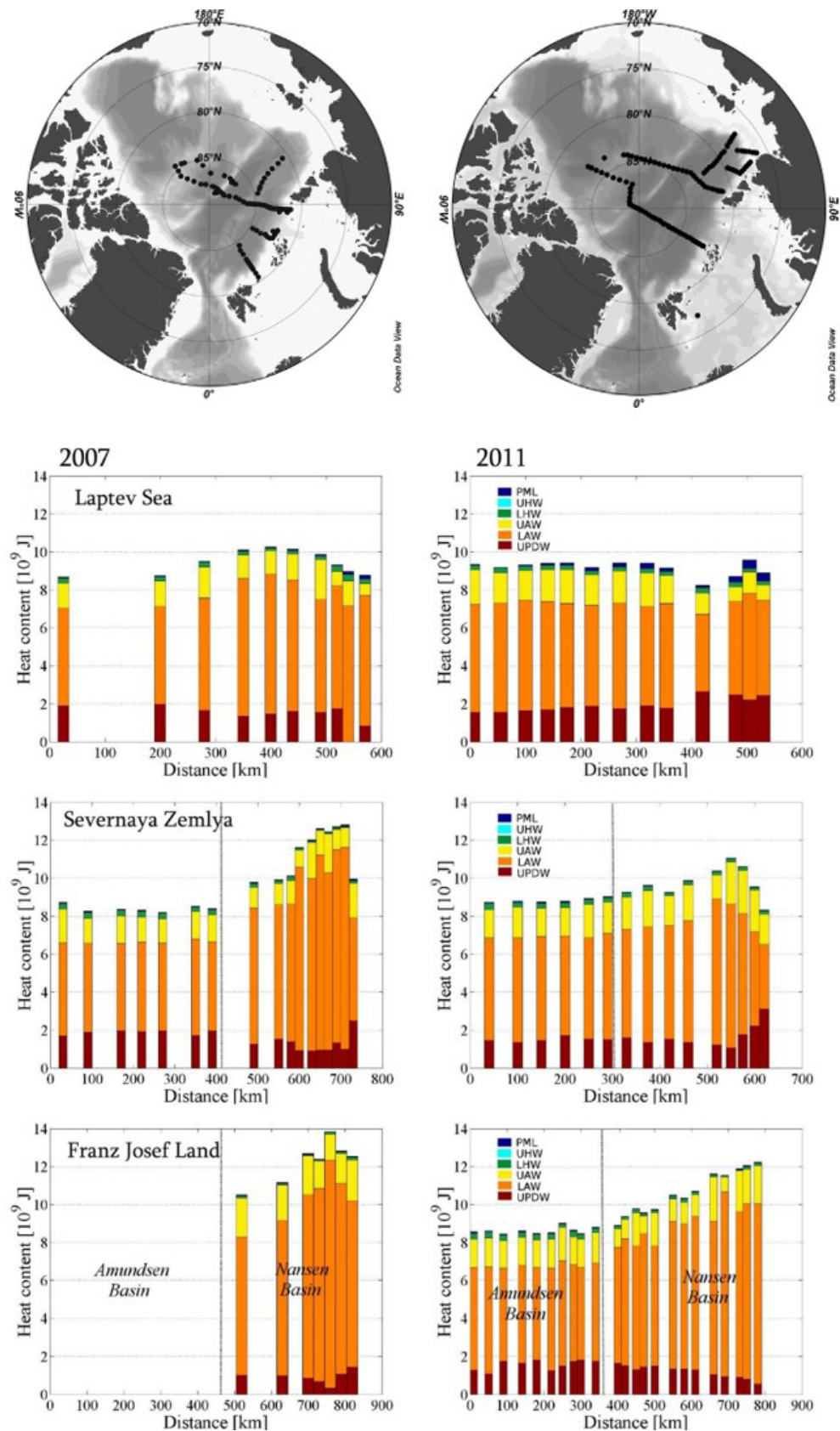


Figure 9.

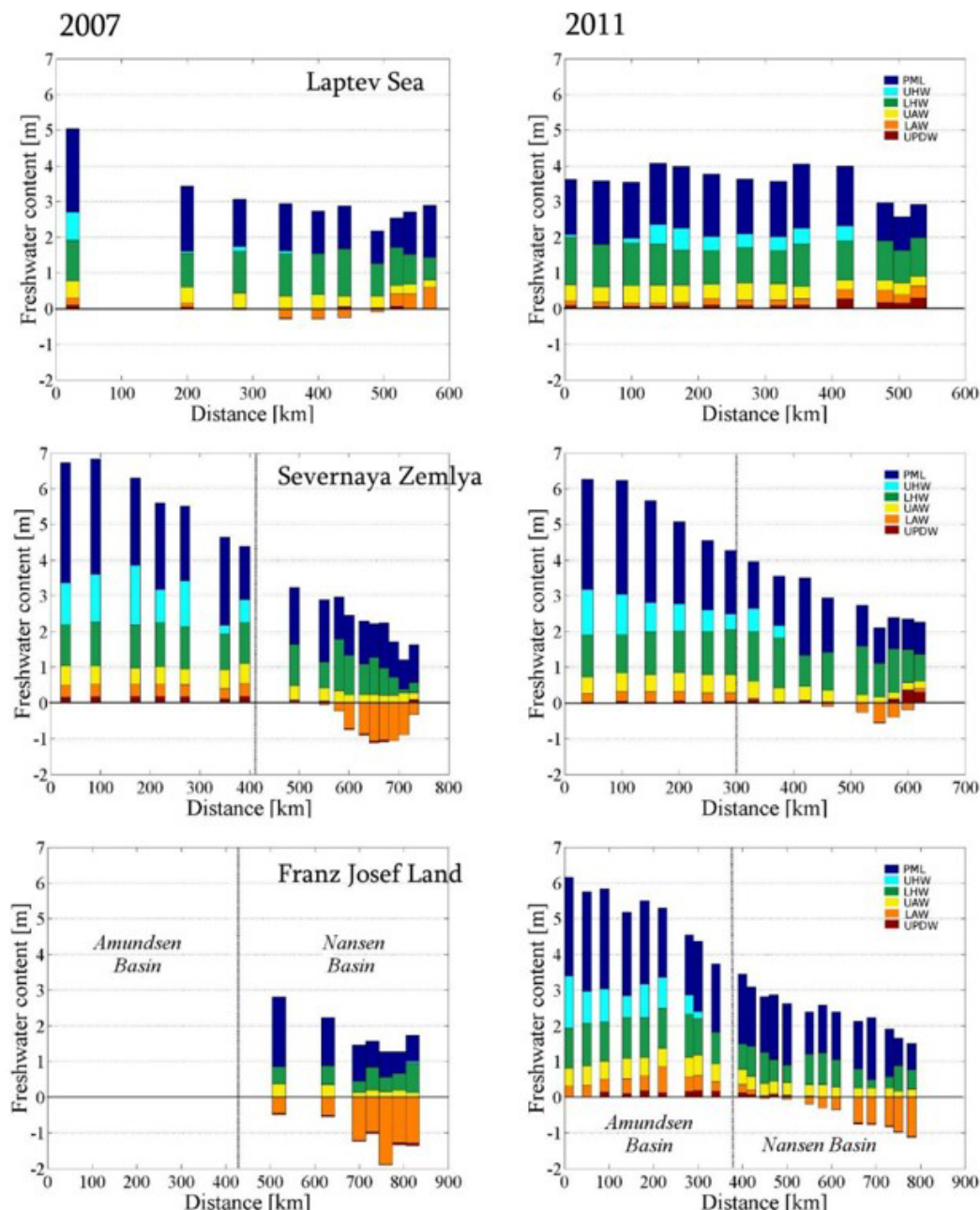


Figure 9. Heat content (central panels) and freshwater content (lower panel) in the upper 1000 m of the water column on stations taken on three sections in the Eurasian Basin in 2007 and almost repeated in 2011. Section positions are shown in the top panel. The heights of the columns indicate the total heat content relative to the freezing point. The freshwater content is shown relative to 34.9 but here the height of the water column only shows the total freshwater content if the freshwater content is positive in all water masses. Too saline water, like the Atlantic water, has negative freshwater content and has to be subtracted. The different waters are; the Polar Mixed Layer (PML), located above the upper temperature minimum indicating winter convection, upper halocline water (UHW), found below the upper temperature minimum and with $S < 34$, lower halocline water (LHW) defined by $\theta < 0^\circ\text{C}$ and $34 < S$, Atlantic water I (AWI) $0^\circ\text{C} < \theta$, and above the temperature maximum. Atlantic water II (AWII) $0^\circ\text{C} < \theta$, but below the temperature maximum, upper Polar Deep Water $\theta < 0^\circ\text{C}$, below AWII and above 1000 db. Especially the increase in freshwater content in the Atlantic layer on the eastern sections and the input of freshwater to the upper layer across the Laptev Sea shelf break are clearly seen (figure from Rudels et al., 2015).

interleaving into a thermohaline step structure with thick homogeneous layers separated by thin diffusive interfaces with large temperature steps and small stability ratios (Rudels, 2021). Such thick diffusive layers have been observed in the eastern Nansen Basin above the temperature maximum and they would contribute to a vertical rather than a lateral heat transport (Polyakov et al., 2012).

The temperature of the Atlantic water in the Nansen Basin was in 1990 observed to have increased by about 1 °C compared to the climatology (Quadfasel et al., 1991). During the 1990s more changes in the Arctic Ocean water masses were observed. Higher Atlantic water temperatures were also seen in the Makarov Basin and the Pacific water that previously supplied the upper layer in the Makarov Basin had retreated from the Lomonosov Ridge to the Mendeleyev Ridge (Carmack et al., 1995; McLaughlin et al., 1996). This warming trend eventually reached and extended across the Beaufort Gyre of the Amerasian Basin in early 2000's (McLaughlin et al., 2009).

Also, the runoff normally advected from the Laptev Sea to the Eurasian Basin was absent, and the more saline melt water layer extended to the sea surface also in the Amundsen Basin, bringing the Atlantic water closer to the sea surface and the ice cover. By contrast, in the Canada Basin the upper layer had become thicker and less saline, while the Atlantic water was pressed down and located at deeper levels (Morrison et al., 1998).

These changes in the Arctic Ocean water column were likely due to changes in the atmospheric circulation. During the 1990s the North Atlantic Oscillation Index was positive and high, showing that the Icelandic low pressure system was strong and extended north into the Barents Sea, perhaps farther. This not only brought more Atlantic water into the Arctic Mediterranean, it also forced the runoff from the large Siberian rivers to leave the shelves farther to the east, into the Makarov and Canada basins (Maslowski et al., 2000).

Several later pulses of warm Atlantic water have entered through Fram Strait and been followed as they moved around in the different gyres in the Arctic Ocean (Polyakov et al., 2005, 2011). Changes have, however, also occurred in the Barents Sea branch. In the 1980s the Barents Sea branch water that entered the Arctic Ocean was colder, less saline and denser and showed up as a salinity minimum between 800 and 1200 m in the Eurasian Basin (Rudels et al., 2013, see also Fig. 8). In the 2000s, by contrast, the water that entered and passed over the Barents Sea to the Arctic Ocean was warmer, more saline and denser and have instead been identified in the Eurasian Basin as a salinity maximum around 2000 m (Rudels, 2021).

The densest water created by ice formation and brine rejection on the shallow areas west and north of Novaya Zemlya, however, does not show up clearly in the Arctic Ocean water column. Its presence has been deduced from an increase in temperature in thin saline bottom layers at the slope (Rudels et al., 2000). The high salinity is due to brine

rejection and accumulation of saline water at the shelves, and the temperature increase is caused by entrainment of intermediate warm Atlantic water as the plumes sink into the deep. This process has, however, been followed in more detail in the western Barents Sea (Fig. 10), where brine enriched dense outflows from Storfjorden in southern Svalbard sink into the deep Fram Strait, entraining warm Atlantic water (Quadfasel et al., 1988).

There are two different conceptual models for descending plumes. Shaving plumes (Aagaard et al., 1985), where the interfacial stress on the upper part of the plume causes the plume to gradually lose mass to the surrounding water as it sinks into the deep. The second model pictures thin plumes passing through and entraining water from the intermediate layers (Rudels, 1986a). The turbulence is generated by bottom stress and to accommodate the increased mass transport the plumes do not grow vertically but laterally and their small thickness is retained.

Shaving plumes are initially fairly thick and carry a substantial amount of dense water from the shelf into the deep and redistribute it to the ambient water column as they sink. This is a picture that would fit the denser water descending down the St. Anna Trough, and perhaps also the dense overflow plumes crossing the Greenland–Scotland Ridge as well as descending Antarctic slopes (Carmack, 1973). Entraining plumes, by contrast, would be thin and carry little but highly saline water from the shelf and instead incorporate intermediate water from the bypassed layers and bring it into the deep. The Atlantic water flowing along the slope thus becomes transported downwards and transformed into deeper and denser water masses. The increase of temperature in the bottom layer (Fig. 10) indicates that the boundary plume, initially at the freezing point, has entrained warm Atlantic water during its descent and the higher salinity relative to the overlying water column shows that some of the initial salinity excess formed at the shelf still remains. The profiles and θS curves seen in Fig. 10 thus suggest entraining plumes.

7.2 A possible wind driven circulation

The two Atlantic inflows partly cross the Lomonosov Ridge and enter the Amerasian Basin, partly recirculate in the Eurasian Basin and return directly to Fram Strait, mostly along prominent bathymetric features such as the Lomonosov Ridge and the Gakkel Ridge. The forcing of the Atlantic water circulation in the Arctic Mediterranean deep basins is not well established, but many authors have presented unified conceptualizations of the circulation of Atlantic waters as a boundary-trapped contiguous flow above the continental slope (cf. Aagaard, 1989; Rudels et al., 1994; McLaughlin et al., 2002; Aksenov et al., 2011; Bluhm et al., 2020). Here the approach taken by Nøst and Isachsen (2003) will be examined.

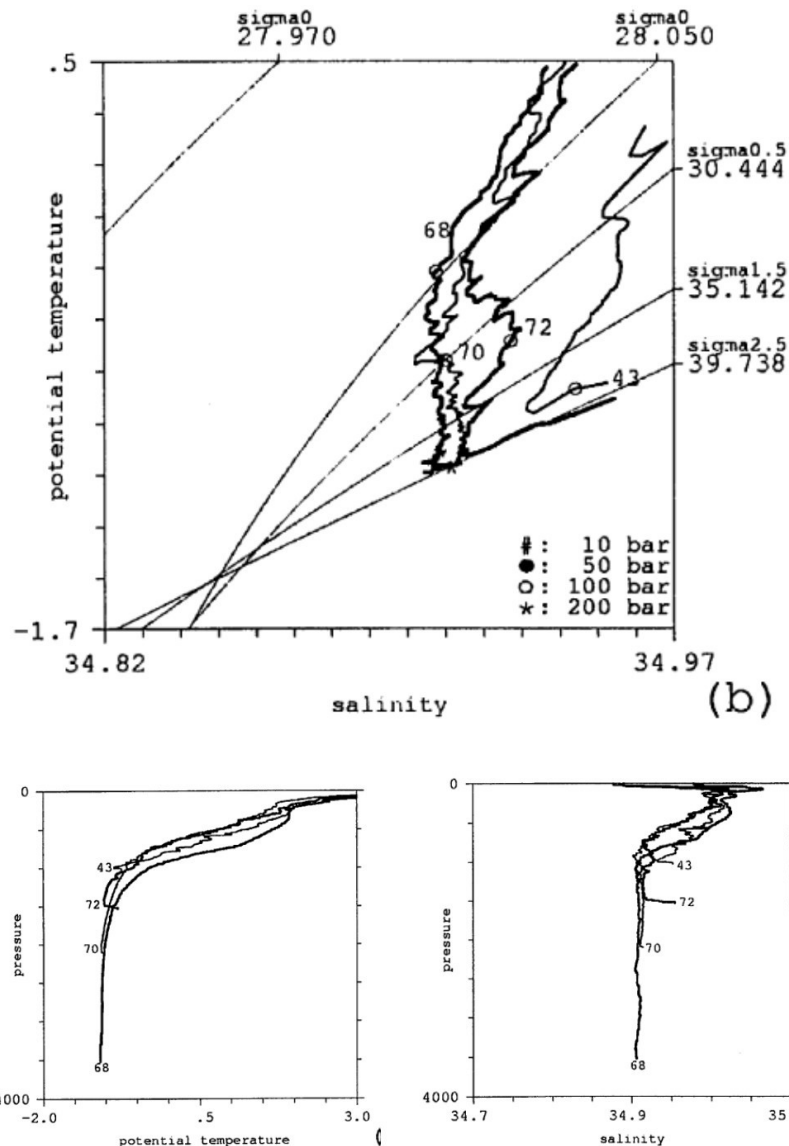


Figure 10. Potential temperature and salinity profiles and θS curves observed from FS Valdivia west of Svalbard in 1986. The increase in salinity and temperature in the deepest layer indicates brine enriched water from Storfjorden that has entrained intermediate warm Atlantic water during its descent (adapted from Rudels et al., 1999b).

Nøst and Isachsen (2003) assumed that the circulation is wind driven and computed the vorticity input at the surface from the NCEP/NCAR reanalysis data (Kalnay et al., 1996, 1997) and the thermal wind transports from the observed seasonally averaged hydrographic fields. They found that these did not balance and a Sverdrup circulation, where the added wind torque drives a meridional transport, is not present in the Arctic Mediterranean Sea. The vorticity added at the sea surface must then be removed by nonzero bottom velocities and bottom friction.

Nøst and Isachsen (2003) suggested either quadratic or linear bottom friction laws to represent the bottom torque. The applied meteorological forcing corresponds to a vertical ve-

locity of about 10^{-6} m s^{-1} , and the vertical velocity generated by the bottom torque must be of the same order. The bathymetry of the Arctic Mediterranean consists of several basins that display closed topographic contours H and also closed geostrophic contours f/H , H being the depth and f the Coriolis parameter. Assuming that the bottom circulation follows the geostrophic contours and that the width of the bottom streams is equal to the width of the continental slopes, the bottom velocities can be estimated to 10^{-1} m s^{-1} , which is close to the observed bottom velocities (Newton and Sotirin, 1997; Nøst and Isachsen, 2003). The picture is then of an almost barotropic deep circulation in the Arctic

Mediterranean that largely follows the f/H contours and that the wind forcing is balanced by the bottom torque.

The neglect of thermodynamics, except that present in the observed and used density field, prevents the model from addressing the changes in the stratification that take place in the Arctic Mediterranean. Aaboe and Nøst (2008) tried to remedy this limitation by introducing a variation in bottom density along the f/H contours. The bottom density is highest in the Nordic Seas and lower in the Arctic Ocean and especially in the Canada Basin. This generates a cyclonic flow in the Nordic Seas, which weakens and eventually changes to anticyclonic in the Arctic Ocean. This change can be explained by a shift from a strong barotropic circulation in the Nordic Sea to a larger contribution of the baroclinic part of the circulation (Aaboe and Nøst, 2008). No thermodynamics is included in the model that can explain the bottom density changes, but one possibility could be that the input of shelf water and the Pacific inflow at the surface pushes the boundary current to deeper levels and replaces the water column at the chosen f/H contour with a less dense one from higher up on the slope.

Nøst et al. (2008) made a theoretical and experimental study of two connected basins with sloping bottom and with one basin driven by either cyclonic or anticyclonic forcing. They found that cyclonic forcing at one basin also made the second basin exhibit a cyclonic circulation. By contrast, a weak anticyclonic forcing in one basin creates an anticyclonic circulation in the second basin. However, when the anticyclonic forcing increases the circulation in the second basin shifts to cyclonic. It thus appears that a cyclonic circulation is more stable than an anticyclonic one. Another interesting detail is that cyclonic forcing in one basin can also drive a cyclonic circulation in an adjacent basin. This implies that an atmospheric forcing in the Nordic Seas could drive a cyclonic circulation along the geostrophic contours also in the Arctic Ocean. If the cyclonic wind forcing in the Subpolar gyre can drive the Atlantic inflow to the Arctic Mediterranean is then another interesting possibility.

8 The conditions in the Amerasian Basin

8.1 The upper layers and the ice cover

East of the Laptev Sea and the New Siberian Islands the nature of the circulation changes. It is no longer dominated by the two Atlantic inflows but by freshwater input from the low salinity Pacific water and from rivers and continental runoff. The inflow through Bering Strait has since the beginning of this century been monitored by direct current measurements and the estimates indicate an increase in volume and freshwater transports as well as in heat transport (Woodgate and Aagaard, 2005; Woodgate et al., 2012, 2015; Itoh et al., 2012, 2026). The increased heat and freshwater transports might largely be due to the observations also of the mainly sea-

sonal Alaskan Coastal Current that carries freshwater from the Yukon river discharging into the Bering Sea (Woodgate and Aagaard, 2005). The mean yearly volume transports are presently estimated to 1.2 Sv and about one third the freshwater input (relative to 34.80) to the Arctic Ocean takes place through Bering Strait (Dickson et al., 2007).

The resulting freshwater storage confines the atmospheric forcing and the oceanic response to the upper part of the water column. The mainly wind driven circulation is centered below the atmospheric high pressure cell above the Beaufort Sea in the Canada Basin. The anticyclonic wind field forces the less dense surface waters towards the center of the Beaufort Sea, leading to a higher sea level and to an anticyclonic geostrophic circulation, the Beaufort Gyre, in the upper few hundred meters.

In the Eurasian Basin, by contrast, the low pressure cell extending from Iceland into the Barents Sea creates a cyclonic circulation that drives Atlantic water from the Norwegian Sea into the Arctic Ocean and forces much of the continental runoff from Eurasia into the Amerasian Basin.

The two gyre systems are brought together and interact in the Transpolar Drift, where the Siberian branch originating in the Laptev Sea flows towards Greenland and Fram Strait in the Amundsen Basin parallel to the northward flowing part of the Beaufort Gyre. Waters are exchanged between the two gyres and the denser Siberian branch intrudes into and supplies water to the deeper levels of the Beaufort Gyre (Morrison et al., 2012). Both streams transport sea ice and especially the Laptev Sea has been considered an important formation site for the sea ice exported from the Arctic Ocean.

The existence of the Transpolar drift was demonstrated by the drift of *Fram* 1893–1896. Roald Amundsen tried to repeat the drift with *Maud*, 1918–1925, but by starting from the East Siberian Sea farther to the east, he hoped to have a greater possibility to drift over the North Pole. This plan did not work. *Maud* remained in the East Siberian Sea trapped by the ice and did not pass beyond the shelf break. The main source of the Transpolar drift thus appears to be located west of the New Siberian Islands, while the East Siberian Sea in contrast to the Laptev Sea receives a considerable amount of sea ice from the central basins and more acts as an ice sink than an ice source for the inner Arctic Ocean. This is partly confirmed by Zacharov (1976), who estimated the ice export from Siberian shelves and concluded that the Laptev Sea ice export was about ten times as large as that from the East Siberian Sea.

The ice cover in the East Siberian Sea has diminished drastically in recent years and has now changed to seasonal from being present almost the year around. This is but one indication of the changes taking place in the Arctic Ocean ice cover. Such changes may even affect structures deep within the water column (Bretones et al., 2022). The decline in sea ice extent has since the beginning of the satellite observations in the 1970s been followed almost in real time, but information about sea ice thickness is more difficult to obtain. Not until

upward-looking sonar observations were made from the submarine USS Pargo in 1993 could sea ice thickness measurements be compared with those made about 30 years earlier. The thickness of the Arctic Ocean ice cover had decreased by 40 %, from an average thickness of about 3 to 1.8 m (Rothrock et al., 1999).

The thickness reduction continued and accelerated in the following decades (Kwok et al., 2009) and a major shift in the sea ice distribution was observed. The earlier dominance of thick and deformed multiyear ice changed into a stronger presence of thinner first year ice (Kwok et al., 2009; Sumata et al., 2023). This shift appears to have taken place in the first decade of this century (Kwok et al., 2009), and Sumata et al. (2023) relate the change to the strong sea ice export events that occurred in 2005 and 2007. Once the main storage of multiyear ice had been removed the drift speed of the sea ice increased and its residence time decreased, which prevents the buildup of a large volume of multiyear ice in the Arctic Ocean.

The effects of these changes were clearly demonstrated by the ship *Tara*, which in 2008 within the EU-DAMOCLES (Developing Arctic Modeling and Observing Capabilities for Long-term Environmental Studies) program repeated the drifts of *Fram* and *G. Sedov* in only 18 months compared to 3 years for the earlier drifts, showing that thinner ice moves more quickly than the thick ice cover in the past. The reduction in ice cover thickness and extent does not appear to have significantly changed the sea ice export to the Nordic Seas through Fram Strait (Spreen et al., 2009).

The ice grows as it drifts and thinner ice grows more quickly, and the reduced ice thickness is also compensated by the more rapid flow. However, the storage of freshwater as ice in the Arctic Ocean has declined significantly, and the seasonal ice formation manages to supply the ice export but cannot rebuild any large reservoir of especially older, thicker ice within the Arctic Ocean. Some of the more dramatic reductions in the ice cover have been connected with atmospheric circulations and winds favoring a large ice export through Fram Strait as in 2007, the year of the second lowest summer ice extent on record.

The loss of sea ice has not resulted in a reduction of the total freshwater storage in the Arctic Ocean. The total freshwater volume has increased by 4500 km³ and the liquid freshwater that has accumulated in the Beaufort Gyre has increased the freshwater storage there by 5000 km³, between 2000s and the 1980s and 1990s; see also Krishfield et al. (2014). This more than compensates for the about 3500 km³ loss of sea ice (Haine et al., 2015).

8.2 The Beaufort Gyre

The circulation in the Beaufort Gyre is complex (McLaughlin et al., 2009; Timmermans and Toole, 2023; Timmermans et al., 2025; Kelly et al., 2019). The anticyclonic winds collect the low salinity upper water in the center of the gyre.

As a result, the Beaufort Gyre is the major freshwater storage reservoir within the Arctic Ocean (Aagaard and Carmack, 1989). The volume of freshwater within this reservoir is subject to interannual variability and poses a risk of catastrophic discharge into the convective gyre of the northern North Atlantic (Proshutinsky et al., 2009, 2015, 2019). This accumulation is balanced by instabilities in the deeper layers that shed eddies into the surroundings (Manucharyan and Spall, 2016). However, the balance between these two processes is yet not reached, and the present storage of about 20 m of freshwater relative to a salinity of 34.8 could increase to more than 30 m (Manucharyan and Spall, 2016; Manucharyan et al., 2016). Another, additional, process has therefore been suggested, the ice-stress governor, resembling the valve that in the 19th century prevented steam engines from running amok (Meneghello et al., 2018).

The accumulation of low salinity water by the wind in the center of the gyre creates a baroclinic density field that drives an anticyclonic geostrophic circulation around the gyre. If the wind stress on the ice surface is taken up by the internal stresses in the ice cover, the ice cover is no longer forced by the wind. Instead, its motion ceases and it begins to retard the geostrophic circulation of the water below, which in turn reduces the slope of the isopycnals and the accumulation of freshwater. This process should then act together with the instability mechanism to reduce the freshwater storage in the Beaufort Gyre (Meneghello et al., 2018). However, if the ice cover becomes less compact, or perhaps disappears, the ice-stress governor ceases to function and more freshwater would again be collected in the Beaufort Gyre (Doddridge et al., 2019).

With more freshwater accumulated in the Beaufort Gyre the stability increases and the upper layer becomes isolated from the bathymetry. The balance between the input of vorticity at the surface and its removal by bottom stress, as proposed by Nøst and Isachsen (2003), might then disappear and another balance is needed. The possibility of a Sverdrup balance has usually been discarded because of the small β at high latitudes. Yang et al. (2016) examined the situation for a circular basin in high latitudes. They found that in the case of no boundaries the effects of the difference in β between latitudes was less pronounced compared to the case when a western boundary was artificially introduced. They concluded that the vorticity input from the anticyclonic wind stress curl would in the gyre be balanced by the time mean divergence of the eddy vorticity flux and along the western boundary by a frictional torque.

No solid western boundary is, however, present near the Beaufort Gyre, but there is a northward flowing boundary current that is fed both from the east and from the west, the Transpolar drift. The Beaufort Gyre is a unique feature in the world ocean. It is forced by anticyclonic winds, creating the density stratification and the gyre circulation. There should be an, albeit weak, southward Sverdrup transport, which could continue south through the Canadian Arc-

tic Archipelago but also partly could feed an inertial current flowing northward on the western side of the gyre. The increase in planetary vorticity would then be balanced by an increase in negative relative vorticity in the boundary current. The flow would intensify as long as f increases, but when the highest latitude is reached, f starts to decrease and the boundary current begins to diverge, reducing its relative vorticity. This is different from mid latitudes, where a purely inertial boundary current cannot exist because the boundary current must shed its relative vorticity, when it leaves the boundary, which only can be accommodated by friction (Valis, 2017).

8.3 Vertical exchanges, diffusive interfaces and the circulation in the Atlantic layer

In the Amerasian Basin the input of low salinity water from the shelves, and the Pacific inflow that create a temperature maximum around 75 m in summer and form a temperature minimum between 200 and 300 m in winter, further isolate the Atlantic water from the energy input at the sea surface. Variability in the inflow of summer water, which lies immediately below the mixed layer, has been observed to influence the ice cover (Shimada et al., 2006). The underlying temperature minimum also involves an increase of salinity, suggesting that in winter ice has formed in the Bering and Chukchi seas and that the bottom water salinity has increased by brine rejection. This is in agreement with the mechanism for creating the cold Arctic Ocean halocline proposed by Aagaard et al. (1981), which isolates the deeper layers from the surface processes. The cold, saline water also has high nutrient content, indicating nutrient regeneration from the bottom sediments (Jones and Anderson, 1986).

The thermocline between the Pacific winter water and the temperature maximum of the Atlantic layer displays step structures and homogenous layers, suggesting that heat is transported upwards by double diffusion. The individual diffusive layers, about 5 m thick, appear to be coherent over 800 km, giving a horizontal to vertical ratio of about 200 000 to 1 (Timmermans et al., 2008). However, over the same distance large temperature differences are present in both the upper and the lower part of the thermocline. The vertical diffusive heat flux reduces the temperature of the Atlantic core, but the lost heat becomes trapped in and warms the temperature minimum of the Pacific winter water. The temperature in the central part of the thermocline, however, remains more or less constant, suggesting that here the heat fluxes through the lower and upper diffusive interfaces are balanced. This implies that the thermocline and the diffusive layers are advected over the basin on time scales of several years. The layers are coherent not only in space but also in time.

The stability ratios $\beta S/\alpha T$ at the interfaces are large, 3–4. This implies that parcels convecting from the interfaces can homogenize the existing layers but could not have created the initial layers. The layers, or embryos of the steps and layers

might then be present already as the Atlantic water enters the Amerasian Basin and the larger number of layers and interfaces compared to the Eurasian Basin could be due to a decreased buoyancy anomaly of the rising and sinking plumes. They are then no longer able to stir and homogenize the entire layer but only the part close to the interfaces, and new interfaces are created at mid-depth.

The Atlantic water that returns from the Amerasian Basin lies too deep to exit through the Canadian Arctic Archipelago. Instead, it appears, at least partly, to leave the North American continental shelf and slope and move northward as a boundary current along the Lomonosov Ridge. There it passes through gaps in the ridge from the Makarov to the Amundsen Basin and then returns south towards Greenland (Björk et al., 2010). Some Atlantic water may also cross the Lomonosov Ridge at the continental slope directly to the Amundsen Basin.

8.4 The deep waters in the Arctic Ocean

In the mid-20th Century the commonly accepted view was that the deep and bottom water source in the Arctic Mediterranean was located in the Greenland Sea and that cold, dense deep water penetrated into the Arctic Ocean through Fram Strait and its temperature increased with distance from the source (Wüst, 1941). However, on the US expedition Ski-jump 1951 and 1952 L. V. Worthington noticed that the deep water temperatures below about 1600 m in western Arctic Ocean, away from Fram Strait, remained almost constant around -0.5°C , significantly higher than in the Greenland Sea. Worthington suggested that the higher temperature was caused by a submarine ridge that prevented the deepest and coldest water from spreading to the entire Arctic Ocean (Worthington, 1953). Unknown to Worthington this ridge, the Lomonosov Ridge, had been discovered by the Soviet airborne high-latitude expeditions in 1948.

It was later noticed that not only the deep temperatures but also the deep salinities were higher in the Amerasian Basin beyond the Lomonosov Ridge than in the Eurasian Basin (Aagaard, 1980). This implies that the deep Arctic Ocean is not stagnant and that water mass renewals must take place. The situation was akin to the one Nansen had encountered with the erroneous, too high, salinity measurements from *Fram*. Nansen had initially suggested that these salinities could be due to brine enrichment of the bottom waters on the shallow shelves and a subsequent sinking of the dense shelf water down the continental slope (Nansen, 1906). An idea that he later abandoned, when he found that the deep salinities measured north of Svalbard agreed with those observed in the Greenland Sea (Nansen, 1915). Now the Arctic Ocean was suddenly recognized as an additional active source for the deep water in the Arctic Mediterranean and Aagaard et al. (1985) and Rudels (1986a) proposed a connected system, where the Greenland Sea supplies the colder, less saline end member and the Arctic Ocean the

saline, warmer end member and the two sources communicate through Fram Strait.

The water columns in the Eurasian Basin and in the Amerasian Basin differ. The Eurasian Basin is warmer and more saline in its upper part while the Amerasian Basin is warmer and more saline in the deep. The changeover occurs at different depth levels for temperature and salinity. The Amerasian Basin becomes warmer than the Eurasian Basin below 900 m, well above the sill depth of the Lomonosov Ridge, but it becomes more saline first at 1400 m, closer to the sill depth, but still in a depth range, where the different basins communicate with each other.

These differences must be due to slope convection and the entrainment into the sinking boundary plumes. The colder deep water in the Eurasian Basin could be explained by the entrainment of much colder ambient water as the plumes sink through the deeper part, while the more saline characteristics of the Amerasian Basin may be caused by the less dense upper part of the water column, which requires the initial shelf contributions to be much more saline to reach the deepest layers than those descending from the Eurasian shelves.

Below about 1500 to 2000 m the deep waters in the different basins communicate less freely and they have acquired their different physical, chemical and biological characteristics (Bluhm et al., 2015). The deep water columns in the Nansen, Amundsen and Canada basins are similar. The temperature decreases and the salinity increases until about 1000 m from the bottom, where a temperature minimum is encountered. Below this level the temperature increases slightly and the salinity continues to rise until a homogenous, 600 to 800 m thick bottom layer is encountered. The temperature increase and the homogenous bottom water have been explained by geothermal heating, which could homogenize the bottom layer by convection from below and perhaps eventually remove the temperature minimum (Timmermans et al., 2003; Björk and Winsor, 2006; Carmack et al., 2012; Arroyo et al., 2025). The existence of a temperature minimum in the three basins is not obvious. In the Canada Basin it can be explained by advection of deep water from the Makarov Basin across the Alpha and Mendeleyev ridges, but no such source can be identified in the Nansen and Amundsen basins.

The initial temperature of the shelf water is at freezing and practically constant while its initial salinity depends on the amount of released brine. The advective water that is entrained into the descending plumes might, however, vary in time. A colder, and denser, advective water column would make the descending plumes colder and denser and the coldest plumes would penetrate the deepest. Combined with geothermal heating of the bottom layer formed by the most saline plumes a deep temperature minimum could then evolve.

The Makarov Basin is different. No deep temperature minimum is present and the salinity increase with depth stops about 1000 m from the bottom, while the temperature con-

tinues to decrease until 500 m above the bottom. Below this level a homogenous bottom layer is found. This feature was explained by Jones et al. (1995) as a spillover across the Lomonosov Ridge of colder, less saline deep water from Amundsen Basin, which could sink to the bottom, aided by the higher compressibility of colder water. However, when the deepest sill at the Lomonosov Ridge was reached and studied by IB *Oden* in 2005, no such overflow was found. Instead, warm and saline deep water from the Makarov Basin was observed entering the Amundsen Basin (Björk et al., 2007). If there is a spillover from Amundsen Basin to the Makarov Basin, it must be intermittent.

9 Fram Strait

The downstream conditions of the inflow to the Arctic Ocean through Fram Strait and the upstream conditions of the return flows to the strait are shown in the northern section in Fig. 11. The inflow involves not only Atlantic water, which mainly enters over the shelf north of Svalbard, but also denser, deeper lying waters, which cannot pass directly eastward and have to circulate along the Nansen Ridge and the northern slope of the Yermak Plateau before they can join the shelf inflow at the continental slope east of Svalbard. Farther north waters returning towards Fram Strait in the northern Nansen Basin, over the Gakkel Ridge and in the Amundsen Basin dominate. Both the Greenland Sea Deep Waters and the Eurasian Basin Deep Waters above the sill of Fram Strait have warmed in recent years, such that the Greenland Sea is now a source of warming to the interior Arctic Ocean (Karam et al., 2024; Song et al., 2025).

The Arctic Atlantic water returning in the Eurasian Basin is warmer and more saline than the Arctic Atlantic water that has circulated through the Amerasian Basin and approaches Fram Strait along the Greenland slope. The water that enters the Arctic Ocean over the Barents Sea is identified as a colder, less saline layer at 1000 m depth, but it is difficult to separate the intermediate waters recirculating in the Eurasian Basin from those that have passed through the Amerasian Basin, except by location. The deep water from the Amerasian Basin shows up as a salinity maximum around 2000 m at the Greenland slope, and the colder and denser Eurasian Basin deep water fills the deeper layers on the northern section (Fig. 11).

On the southern section, taken along 78° 50' N in Fram Strait (Fig. 11), there is an inflow of mainly Atlantic water in the West Spitsbergen Current but also of intermediate and deep waters from the Nordic Seas. The deep waters in general are slightly less saline in Fram Strait than farther north in the Nansen Basin. The central part of the strait exhibits a strong recirculation of Atlantic water towards the west. The recirculation is largely carried by barotropic and baroclinic eddies that appear to move westward towards Greenland. The main southward flow from the Arctic Ocean takes place in

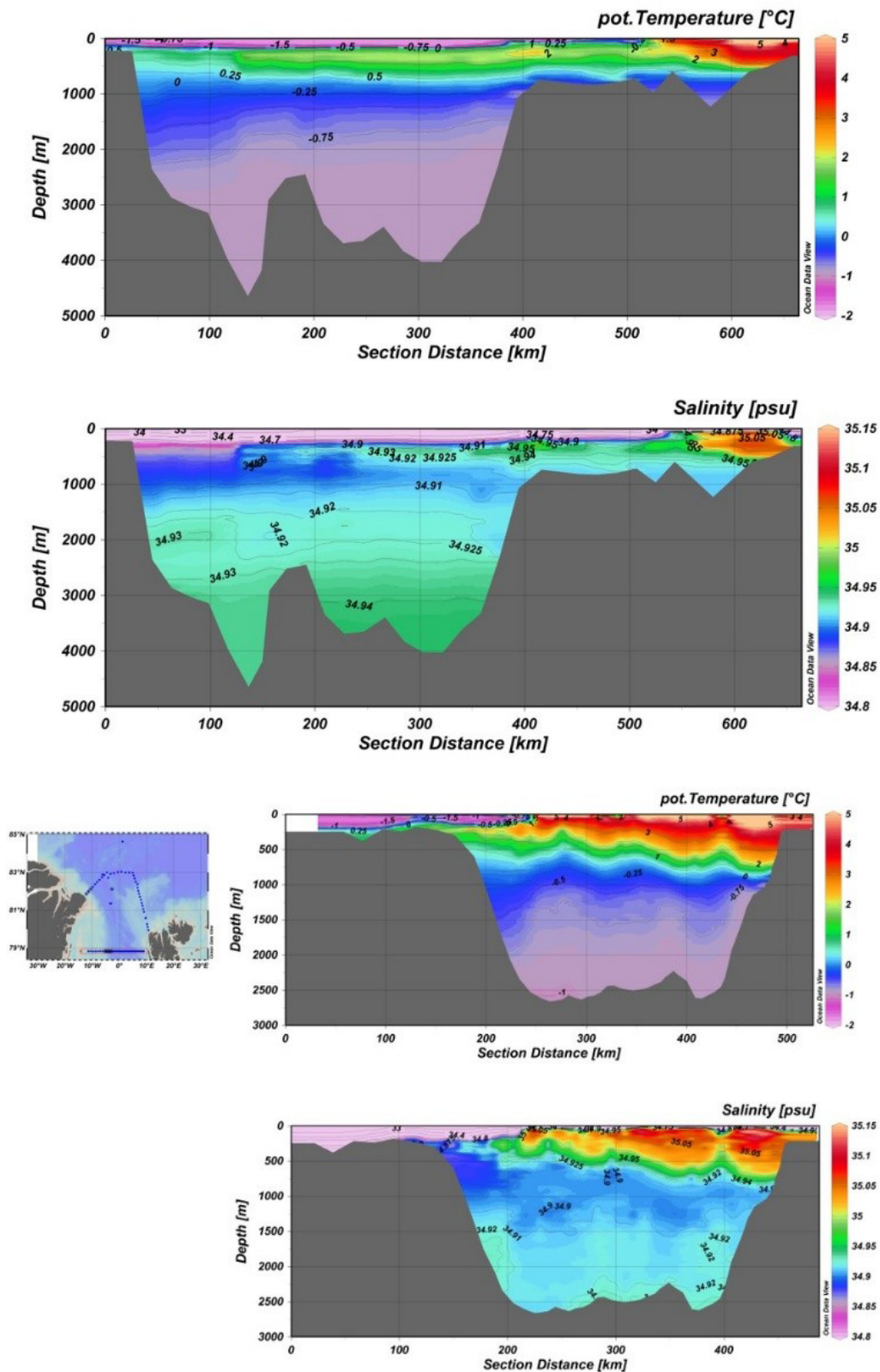


Figure 11. Two potential temperature and salinity sections, one taken from Svalbard along the Nansen Ridge and over the Yermak plateau into the Nansen Basin and then westward to Greenland and the second along the sill in Fram Strait showing the exchanges between the Arctic Ocean and the Nordic Seas (sections taken by FS *Polarstern* 2004).

the west in the East Greenland Current. The recirculation in Fram Strait is strong and only a narrow space is available for the outflowing water between the recirculating Atlantic water and the Greenland continental slope.

The East Greenland Current is more baroclinic than the West Spitsbergen Current and carries several different water masses. The upper layer is dominated by low salinity waters comprising river runoff, occasional contributions from the Pacific inflow, and halocline water created from ice melting on the warm Atlantic water entering north of Svalbard. The cold, low salinity water flows above warm recirculating Atlantic water that penetrates into the East Greenland Current above the transformed Arctic Atlantic water that is returning along the Greenland continental slope. The warmest Arctic Atlantic water derives from the Fram Strait branch that mainly remains in the Eurasian Basin while the colder contributions are supplied by the Barents Sea inflow branch that circulates around the Makarov and Canada basins. The Barents Sea inflow is transformed into an intermediate water mass with distinct doubly stable θS characteristics, salinity increasing and temperature decreasing with depth.

At the western slope two outflows of the Arctic Ocean deep waters can be identified. A warm, saline core centered around 1800 m depth, which ultimately derives from the Am-erasian Basin, and a colder core with similar salinity located at sill depth, indicating an outflow of Eurasian Basin deep water. The comparatively warm and saline characteristics of the Arctic Ocean deep waters indicate that they are created by slope convection as shelf water made dense by ice formation and brine rejection sinks down the slope entraining ambient warmer water. The salinity then decreases and the temperature increases and the plumes continue to sink until their density matches that of the surrounding water column. The deep and bottom water of the Arctic Ocean thus become warmer and more saline than deep water created by open ocean convection. The entrainment into the boundary plumes transports Atlantic water downward in the water column and its density increases as it mixes with the cold, saline shelf outflow. The interaction with sea ice formation and sea ice melting thus separates the inflowing Atlantic water into both less dense and denser waters.

Fram Strait is the only deep passage to the Arctic Ocean and most of the entering Atlantic water must exit through Fram Strait. The low salinity waters can, however, leave the Arctic Ocean either through Fram Strait or through the narrow and shallow passages of the Canadian Arctic Archipelago. The largest runoff enters the Kara and Laptev sea shelves and since the shelf outflow from the Laptev Sea to the Amundsen Basin has salinity between 32 and 33 the runoff must mix with an advective saline water mass, and the only one available on these shelves is the water of the Norwegian Coastal current that enters the Barents Sea and continues eastward to the Kara Sea and Laptev Sea shelves.

The Chukchi shelf is flooded from behind by the Pacific inflow, which also supplies saline water to the East Siberian

Sea, where, in addition, upwelling of saline water from the Makarov Basin takes place. The melting of sea ice on the entering Atlantic water, either north of Svalbard or in the Barents Sea, creates the densest upper layer water and the densest water that can pass through Nares Strait, the deepest passage in the Canadian Arctic Archipelago (Rudels et al., 2004; Dickson et al., 2007).

The major driving forces behind the exchanges through Fram Strait are not identified with certainty, but the largest transports take place at the slopes, which could indicate either a barotropic wind driven flow along geostrophic contours as suggested by Nøst and Isachsen (2003), but also baroclinic transports of mainly less dense upper layer waters along the coastal boundaries (Werenskiold, 1935; Stigebrandt, 1981), or a combination of both. The sea level in the Arctic Ocean is higher than in the Nordic Seas, which favors a southward flow in the East Greenland Current. It could also obstruct the northward flow in the West Spitsbergen Current and force the westward recirculation in the strait.

The presence of eddies in Fram Strait leads to weak horizontal coherence of the velocities and makes it difficult to determine the transports by direct current measurements. Existing hydrographic sections, by contrast, have a much higher horizontal and vertical resolution which allows for better identification of the transports of different waters. The temporal resolution is of course much less, one section per year compared to one record per hour. Several attempts to compute the geostrophic volume transports through the strait and to estimate the heat and freshwater fluxes have been made. One weakness always present is the choice of the unknown barotropic reference velocities, which to some extent must be subjective. However, this and other difficulties may also offer opportunities to reflect upon what the obtained results really imply and we will discuss in some detail the exercise made by Rudels et al. (2008).

Rudels et al. (2008) noticed that the deep water from the Arctic Ocean is more saline than the deep inflow from the Nordic seas, and that the deep inflow also has to leave the Arctic Ocean as deep water ($28.06 \leq \rho_\theta$). They postulated that 0.4 Sv of deep water ($S = 34.935$) is produced in the Arctic Ocean and that 0.2 Sv of deep water ($S = 34.910$) enter from the Nordic Seas. This implies a net deep outflow M , of $0.4 \times 10^6 \text{ m}^3 \text{ s}^{-1}$, carrying a net salt export S of $13.973 \times 10^9 \text{ kg s}^{-1}$. They applied these constraints on the deep part of the strait and required that the kinetic energy of the added barotropic reference velocities should be a minimum. The barotropic reference velocities extend to the sea surface, but no corrections are applied on the shallow parts of the strait, where no deep water is present. The applied constraints promote a saline deep outflow on the western and the less saline inflow on the eastern side of the strait, in accordance with observations.

Rudels et al. (2008) introduced six different water masses and computed the in- and outflow of these waters in four different parts of the strait; the Greenland shelf and slope,

the western and the eastern central parts of the strait, and the Svalbard shelf and slope for each section. The greatest differences were seen between the Greenland and the Svalbard shelves and slopes, while the water flowing north and south in the central part of the strait were similar, indicating a strong recirculation.

The northward and southward volume transports in Fram Strait are not balanced. This implies that the computed heat and freshwater transports depend upon the choice of reference temperature and reference salinity. Rudels et al. (2008) put the reference salinity equal to the mean inflow salinity and the reference temperature equal to the mean outflow temperature. This implies that the inflow carries no freshwater and the outflow no heat.

Computing the geostrophic transport through 16 hydrographic sections obtained between 1980 and 2005 along the sill in Fram Strait the total in and outflows ranged from 5 Sv to almost 15 Sv with an averaged inflow of 6 Sv and an average outflow close to 9 Sv. This is, as expected, less than the transports obtained by direct current measurements, but not alarmingly so. The mean net outflow, 2.5 Sv, 2.8 Sv if the southward flow on the Greenland shelf is included, is larger than that found from the current meter array.

The transports through the other passages of the Arctic ocean are only in one direction. Inflows take place over the Barents Sea and through Bering Strait and runoff and net precipitation add freshwater. The only other outflows occur through the Canadian Arctic Archipelago. The shelf input and the Pacific inflow are cooled to freezing temperature but the part of the Barents Sea inflow that enters the deeper layers of the basins has mean temperatures around -0.5°C . We assume that the transports through the other passages are known and use the estimates provided by Dickson et al. (2007). It is then possible to determine the amount of heat entering and the amount of freshwater leaving the Arctic Ocean through Fram Strait and also how the advective heat is lost in the Arctic Ocean.

The net volume transport, the mean outflow temperature and the northward heat transport referenced to the outflow temperature were computed for each section. The average values 2.8 Sv, $T_{\text{ref}} = 0.7^{\circ}\text{C}$ and 25 TW obtained from the sections were taken to represent the transports and used to determine how the inflowing heat might be distributed in the Arctic Ocean.

Assuming that 1.2 Sv of the Barents Sea inflow with temperature -0.5°C enter the deeper layers and leave the Arctic Ocean through Fram Strait, this cold water transports $c \times (T_{\text{ref}} - (-0.5)) \times 1.2 \times 10^9 = 5.8 \text{ TW}$ into the Arctic Ocean. $c = 4000 \text{ J kg}^{-1} \text{ K}^{-1}$ is the heat capacity of sea water. The remaining part, 1.6 Sv, of the net outflow is less saline and at the freezing point and exits as upper layer water transporting $c \times (T_{\text{ref}} - (-1.8)) \times 1.6 \times 10^9 = 16 \text{ TW}$ heat northward. The northward heat transport in the East Greenland Current of 22 TW then has to be balanced by outflowing warmer returning and recirculating Atlantic water. This implies that al-

most all 25 TW of the northward heat transport through the strait return south as warmer water in the East Greenland Current.

The remaining 3 TW could melt sea ice north of Svalbard and with an Atlantic inflow temperature of 3.1°C this implies that 0.15 Sv is cooled to freezing temperature and that the fraction $\phi = 2\alpha L(c\beta SA)^{-1}$ (see Sect. 5) or about one third of this heat loss, is used to melt ice and the rest goes to the atmosphere. The amount of melted ice added to the created upper layer is then 0.003 Sv.

Introducing the obtained freshwater and volume transports through Fram Strait in the volume and freshwater balances for the Arctic Ocean compiled by Dickson et al. (2007) Rudels et al. (2008) found that the freshwater fluxes into and out of the Arctic Ocean were almost in balance but the volume transports were unbalanced by 1.17 Sv. This suggests that the barotropic velocities added in Fram Strait are too small. However, since the chosen reference salinity is the mean salinity of the inflowing water, we can add a barotropic correction to the northward transports without affecting the freshwater balance. If a barotropic northward transport of 1.1 Sv is added also the volume fluxes become almost balanced. The northward heat transport, however, then increases.

The difference between the mean inflow temperature, 1.6°C , and the mean outflow temperature 0.7°C is 0.9°C , which increases the heat transport by 4 TW. The net outflow that has to be heated to 0.7°C decreases from 2.8 to 1.7 Sv, implying that only 0.5 Sv of the cold surface water must be warmed to 0.7°C , which requires 5 TW. The heat transport through Fram Strait that enters the Arctic Ocean then becomes 19 TW. If all this heat goes to ice melt by cooling the Atlantic water to the freezing point about 1 Sv of Atlantic water is converted to low salinity upper water. 7 TW go to melting sea ice and 12 TW are lost to the atmosphere. This means that 0.021 Sv of ice is melted and the salinity of the upper layer becomes 34.3.

These values are realistic and agree with other, more recent estimates of the Arctic Ocean heat and freshwater balances (e.g., Beszczynska-Möller et al., 2011; Rudels et al., 2015; Rudels, 2016; Bacon et al., 2022). It appears that the heat transport and the effects of the heat carried by the Atlantic water into and around the Arctic Ocean are less than commonly assumed. The existence of strong warm inflow pulses of Atlantic water could perhaps alter this situation and increase the effects of the Atlantic inflow. However, the persistence of the high temperatures in the Atlantic inflow pulses as they circulate around the Arctic Ocean basins (Polyakov et al., 2005, 2011) rather indicates that the heat remains in the Atlantic layer and the heat is eventually returned to the North Atlantic through Fram Strait.

The discussion above is largely ad hoc and relies strongly on subjective reasoning. However, the inverse modeling of the exchanges between the Arctic Ocean and the world ocean has greatly improved as more observations from the different

gateways have become available. Tsubouchi et al. (2012) estimated the transports through all four passages (Davis Strait, Fram Strait, Barents Sea and Bering Strait) for August 2005 when all gateways were monitored and found an oceanic heat flux into the Arctic Ocean of 154 ± 44 TW and a freshwater export from the Arctic Ocean of 155 ± 65 mSv relative to the mean temperature $1.01^\circ\text{C} \pm 0.16^\circ\text{C}$ and the mean salinity 34.67 ± 0.02 on the total cross section. Since all passages were monitored the total heat and freshwater transports are here independent of reference temperature and salinity. However, the heat and freshwater transports through the individual straits may occasionally appear somewhat odd. The largest freshwater export takes place through Fram Strait but primarily as import of saline Atlantic water in the West Spitsbergen Current, not as export of low salinity water in the East Greenland Current.

To estimate the water transformations taking place in the Arctic Ocean Tsubouchi et al. (2012, 2018), Brown et al. (2025) and Bacon et al. (2022) computed the in- and outflows in different density intervals, and the resulting imbalances had to be balanced by mixing across the density surfaces within the Arctic Ocean. The computations revealed that the density interval encompassing the Atlantic water was not balanced and of the inflowing 6.0 Sv only 3.2 Sv returned in the same density interval as modified Atlantic water. 1.0 Sv had become diluted by freshwater and returned as less dense upper layer water, while 1.8 Sv had been transformed into denser water and returned as intermediate and deep waters (e.g., Bacon et al. 2022). The formation of upper layer water is mainly due to sea ice melting north of Svalbard and the creation of denser water takes place by atmospheric cooling largely in the open water area in the southern Barents Sea.

The observations at the gateways thus reveal transformations of the waters within the Arctic Ocean, processes that are not considered in the circulation model by Nøst and Isachsen (2003). The results above, however, are based mainly on observations, but Dörr et al. (2026) showed that the output from a global NEMO ocean hindcast simulation can describe many of the active processes and how they transform the waters entering the Arctic Ocean. The model was forced with atmospheric and runoff data from the DRAKKAR forcing set and was run from 1958 to 2015 and the outputs from 1979 to 2015 were examined. The computed net in- and outflows through the different passages are close to those observed, there are net outflows through Davis Strait (1.8 Sv) and through Fram Strait (3.0 Sv) and inflows over the Barents Sea (3.5 Sv) and through Bering Strait (1.3 Sv).

The focus is on water mass transformations and the circulation is described both in density space, showing the transformation of inflowing water into denser and less dense waters, similar to the approach used by Brown et al. (2025). However, the transformations are also shown in TS space, following the approach introduced by Gösta Walin (1982). The model results in similar water mass transformations as those deduced from observations in the Arctic Ocean and

discussed in the sections above. One significant difference, however, is that about 3 Sv of Atlantic water are transformed into denser water (Dörr et al., 2026), almost twice the transformation estimated by Brown et al. (2025) and Bacon et al. (2022). The main area of this transformation was also here located to the southern Barents Sea.

10 The Greenland Sea

The Nordic Seas were early identified as a possible source area for the cold, dense deep waters of the oceans. Wyville-Thomson and Carpenter found cold, dense bottom water south of the Faroes, Mohn identified a cold, almost homogeneous water column southeast of Jan Mayen, and Amundsen's observations from *Gjøa* in the Greenland Sea in 1901 showed a cold homogeneous dome of water extending from the bottom almost reaching the surface (Nansen, 1906). However, Nansen (1912) judged that the deep convection observed in the Greenland Sea had little global impact and instead postulated that the bulk of the ocean deep water is formed farther south, in the Irminger Sea southeast of Greenland, or in the Labrador Sea.

In the 1970s this view began to change. Satellite observations showed a polynya in the Weddell Sea ice cover 1977 to 1979 and hydrographic observations confirmed that the polynya was related to a cold, less saline water column indicating recent convection (Gordon, 1978). This stimulated a search for and studies of convection, or chimneys (Killworth, 1979), and in the Greenland Sea satellite observations showed several rapid events of ice formation and subsequent disappearance of sea ice, suggesting sinking of cold brine enriched surface water, which was replaced by upwelled warmer water, melting the ice.

Convection occurs as small-scale short events in winter, usually during bad weather conditions making it difficult to observe and study. In the 1970s field experiments dedicated to convection studies were initiated in known convection sites, in the Mediterranean Sea and in the Labrador Sea, and in the 1980s and 1990s most observational and theoretical work in the Greenland Sea was concentrated on convection and many new techniques were introduced. At the same time the importance of the Greenland Sea convection for the formation of dense deep water that could drive the Atlantic meridional overturning circulation, or the global thermohaline circulation as it was then commonly called, was realized, and perhaps more urgent, what were the consequences, if the convection and deep water formation should weaken. In retrospect, in 2026, the three decades between 1980 and 2010 were decades when the largest known changes in the Greenland Sea convection occurred, and when our understanding of the deep water formation in the Arctic Mediterranean deepened.

In the mid-20th century the principal sign of deep convection, a homogeneous water column extending from the surface

to the bottom, had not been observed in the Greenland Sea and the early observations by Mohn and Amundsen were not fully trusted because the instruments used then were not of the same quality as those deployed during the 1950s. Metcalf (1955) suggested a sloping descent of the dense water that would not be captured by vertical hydrographic casts but this idea was never really accepted. The water column in the central Greenland Sea had structure that appeared to be conserved, even if its deep and bottom waters were renewed. Generally, it consisted of a low salinity upper layer above a deep temperature maximum and further down a salinity maximum below which the temperature and salinity decreased with depth until the homogenous Greenland Sea bottom water was encountered. How could this structure be explained?

Carmack and Aagaard (1973) suggested that the deep and bottom water formation was driven by double-diffusive transports through diffusive interfaces. Atlantic water circulating around the Greenland Sea gyre penetrates towards the center while losing heat upward to the colder surface water but retaining its high salinity. When it finally reaches the center of the gyre, its density has become high enough to renew the deep water and perhaps sink to the bottom. The fact that the convecting water is not in direct contact with the atmosphere could also explain the low oxygen content observed in the Greenland Sea deep water (Carmack and Aagaard, 1973). The importance of double-diffusion was further explored by McDougall (1983), who found that the amount of dense water that could be created by this process corresponded well with existing estimates of deep water formation in the Greenland Sea.

Deep convection areas are usually related to cyclonic circulation that forces the surface water towards the rim and allows the deeper layers at the center to rise as a dome towards the surface. A smaller volume of upper layer water must then be cooled before convection starts and the water column overturns more readily (Killworth, 1979, 1983; Marshall and Schott, 1999). The smaller volume also allows for the possibility that the convection occurs in a filling box mode, where the upper layer density increases in winter, becomes unstable and convects to the bottom, bypassing the intermediate layers (Baines and Turner, 1969). The difficulty with this explanation, in contrast to e. g. brine rejection over the shelves, is that it is not possible to accumulate dense water at the surface. A Rayleigh number based on the ocean depth would directly make the water column unstable, and even with a Rayleigh number based on the penetration depth of the heat anomaly from the surface (Howard, 1966) no large positive anomaly could be created before convection starts.

This fact and satellite observations showing an interplay between the ice tongue Odden and the ice free Nordbukta, suggesting rapid formation and melting of the ice cover in the central Greenland Sea, opened for the possibility that open ocean deep convection could be driven by ice formation (Rudels, 1986a). As the water is cooled to freezing temperature, ice will form and rise to the surface, leaving more saline

water droplets behind. These droplets sink through and their excess salinity diffuses into the ambient upper layer. As the droplets eventually lose their salinity it is instead the ambient upper layer that has become saline and dense with respect to the underlying water and starts to convect. The density and thickness of the unstable layer can be estimated from how long it takes for the salinity anomaly to diffuse, using molecular diffusion coefficients, into the ambient water, and how fast the parcels sink, computed from Stokes' resistance law.

The unstable upper layer then sinks as entraining plumes or thermals into the deep until the plumes reach their neutral density level, where they spread out. Convection, however, continues and the upper part of the water column becomes gradually denser until its density is high enough for the convection to penetrate still deeper. The transformed upper layer then convects into the deep and becomes replaced by warmer water from below that starts to melt the ice and the convection stops. If the cooling persists, the upper layer is again cooled to freezing temperature and ice starts to form, brine is rejected, and another convection cycle starts (Rudels, 1986a; Rudels et al., 1999a). Since it is cold water that convects, it is possible that the thermobaric effect, arising because cold water is more compressible than warm water, further increases the density of the plumes, if the convection reaches sufficiently deep.

This scenario is highly speculative, perhaps even controversial, and field studies were initiated in the Greenland Sea to capture at least some of the involved processes (Quadfasel and Rudels, 1990). During strong cooling and ice formation a sediment trap was deployed 0.5 m below the sea surface to capture the released brine droplets. The brine was collected in sample bottles in initially deflated plastic bags. The sampling interval was 40 min and seven samples were collected. During the time of active ice formation three samples with salinity above the background salinity were collected. To observe the buildup and ejection of the saline top layer a 25 m long drifting tube with temperature and conductivity sensors at 2.5 and 22.5 m was deployed. The experiment only lasted about six hours, but a gradual increase and a sudden lowering of the salinity was observed at the upper instrument but not seen at 22.5 m. This could perhaps indicate a buildup and release of a salinity anomaly (Quadfasel and Rudels, 1990; Rudels and Quadfasel, 1991).

With the advent of the CTD the resolution and accuracy of the salinity and temperature measurements increased tremendously and unexpected details were observed in the Greenland Sea water column. θS structures on stations taken in the Greenland Sea in 1993 by FS *Valdivia* are shown in Fig. 12. The densest water is the Greenland Sea bottom water concentrated in a narrow θS range with temperature and salinity decreasing towards the bottom. Above the bottom water the salinity range increases with increasing temperature and two salinity maxima can be identified. The highest salinities are found over the Greenland continental slope, along the Jan Mayen Ridge and at the Mohn–Knipovich Ridge and

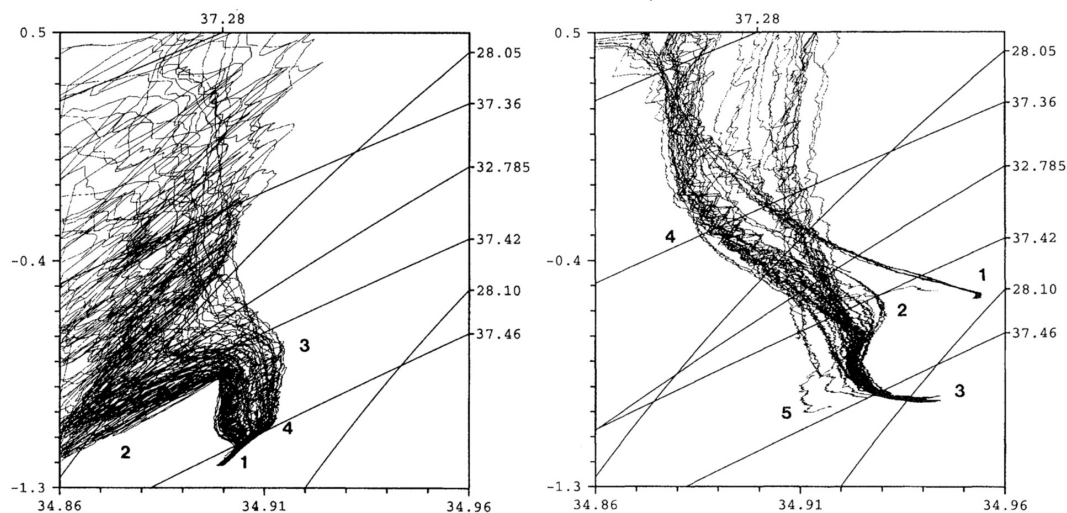


Figure 12. Left panel: blow-up of θS curves from stations taken by FS *Valdivia* in 1993 in the Greenland Sea. The numbers indicate (1) Greenland Sea bottom water, (2) locally formed Arctic intermediate water, (3) intermediate temperature and salinity maxima, (4) deep salinity maximum. Right panel: blow-up of θS curves from stations taken by IB *Oden* in the Arctic Ocean 1991. The numbers indicate (1) the deep and bottom water in the Makarov Basin, (2) the salinity maximum in the Eurasian Basin derived from the Amerasian Basin, (3) the salinity maximum of the Eurasian Basin deep water, (4) stations from the Amundsen Basin showing interleaving, (5) traces of Norwegian Sea deep water north of the Yermak Plateau (from Rudels, 1995).

the lowest at the center of the Greenland Sea gyre. This implies that saline deep waters are advected from north along the Greenland slope and become deflected eastward to circulate around the rim of the Greenland Sea.

The saline deep waters must ultimately arrive from the Arctic Ocean and Fig. 12 also shows θS curves from stations in the Arctic Ocean obtained by the icebreaker *Oden* in 1991. Two saline deep water modes are clearly seen. Warmer and saline deep and bottom water are observed in the Makarov Basin close to the Lomonosov Ridge. Some of this deep water crosses the sill (1800 m) of the Lomonosov Ridge and enters the Amundsen Basin, where it can be identified as a salinity maximum around 1700 m (Anderson et al., 1994; Björk et al., 2007). The deep and bottom waters formed in the Eurasian Basin are colder and denser but slightly less saline. The density of the Eurasian Basin Deep water corresponds well with the deeper salinity maximum observed in the Greenland Sea, while the density of the upper salinity maximum, also forming a deep temperature maximum in the central Greenland Sea, indicates the presence of the Amerasian Basin deep water.

These observations show that the deep waters in the Arctic Mediterranean Sea are supplied by two sources. The open ocean convection area in the Greenland Sea provides the colder, less saline endmember and the Arctic Ocean the saline, warmer endmember. The deep waters communicate through Fram Strait and depending on the regional climate, one or the other source will dominate (Aagaard et al., 1985; Rudels, 1986a).

The salinity maxima are weakest in the central part of the Greenland Sea as is seen in Fig. 13 taken by RV *Lance* 1988, where an eddy exchange between the center and the rim of the Greenland Sea is captured. The exchange is isopycnal and the profiles show the doming of the density field in the central Greenland Sea, shifting the water from the rim upwards and that from the center down to deeper levels at the rim. It is thus likely that the saline Arctic Ocean deep waters that penetrate to the inner part of the Greenland Sea become diluted by the convection taking place there, largely removing their temperature and salinity anomalies.

Figure 14 displays the potential temperature and salinity profiles and the θS curves of two stations taken in the center of the Greenland Sea in 1988, station 65 in February and station 33 in June and a third station, 13, taken in 1993, all from FS *Valdivia*. The February station shows convection deeper than 1000 m and the formation of a homogenous, less saline layer covered by cold, low salinity surface water and located above a more saline deep layer. In June the profiles suggest that convection has continued and reached deeper than 2000 m. The salinity in the upper 1000 m has increased and the salinity in the deeper part has decreased, implying that during late winter the upper layer density has increased and the two layers have been homogenized. The density of the low salinity surface layer has to increase, first by cooling to freezing temperature and if this is not sufficient for convection, ice formation and brine rejection continue to increase the density to above that of the layer below, and the surface layer then convects. The density of the first layer below has then increased sufficiently for it to convect into the second

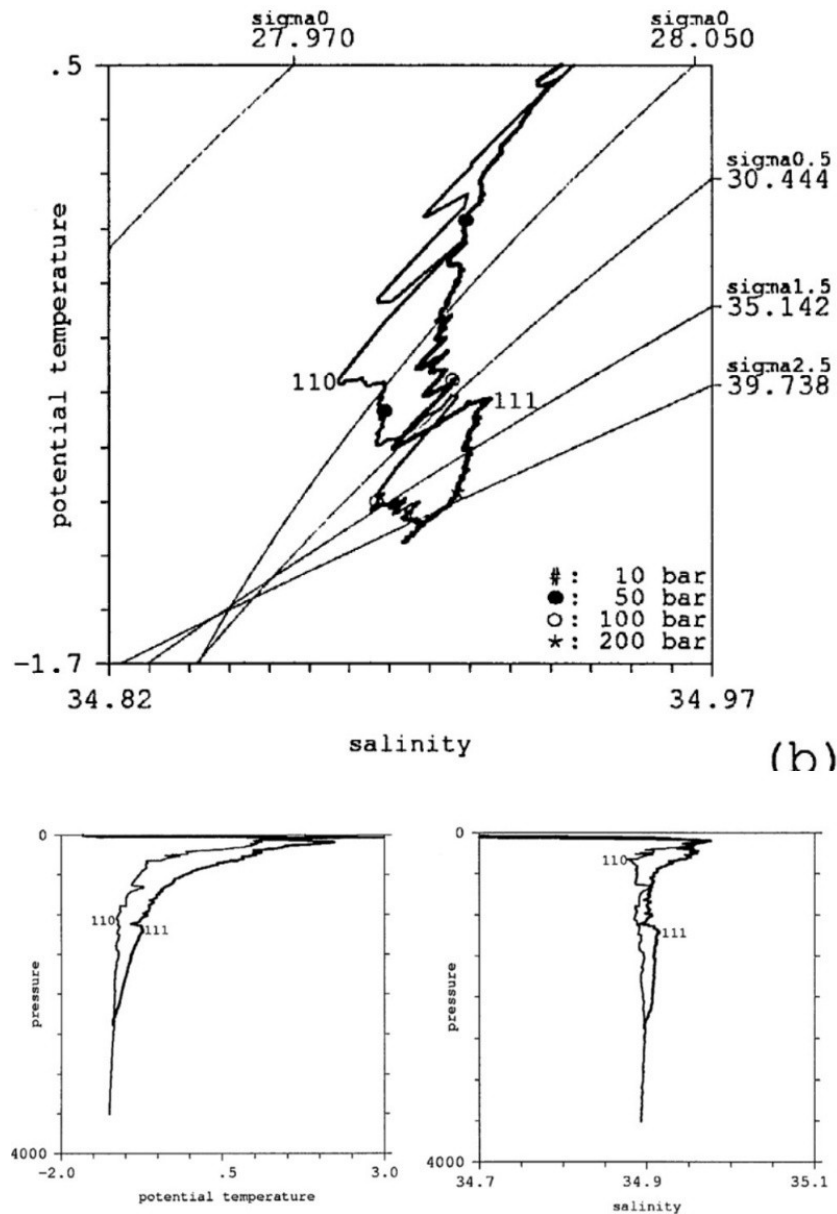


Figure 13. Potential temperature and salinity profiles and θS curves obtained in 1988 from RV *Lance* from stations over the Greenland continental slope (111) and the central Greenland Sea (110). The stations show an isopycnal exchange of eddies between the slope and the central Greenland Sea (adapted from Rudels et al., 1999b).

deep layer below. The two layers then merge, forming a more than 2000 m deep, homogenous water column.

Station 13 in Fig. 14 shows a warmer, more saline water column and deep coinciding temperature and salinity maxima. The convection here extends to the temperature maximum and the convected water forms a temperature minimum above the temperature maximum. The deeper salinity maximum has become more saline and the Greenland Sea bottom water salinity has increased. That the convection is shallower implies that the Arctic Ocean deep water will not be transformed in the central Greenland Sea and the deep salinity

maxima become more prominent. This especially applies to the deeper maximum from the Eurasian Basin.

Figure 15 shows two stations, one from 1982 taken by RV *Hudson* and one *Valdivia* station from 1993. The temperature and salinity of the deep salinity maximum have increased and also the Greenland Sea bottom water has become warmer and more saline. The salinity increase in the bottom water can only be caused by turbulent mixing from above and Meincke et al. (1997) estimated the vertical diffusion coefficient for the deep mixing in the Greenland Sea by assuming a balance between the lateral advection of Eurasian Basin

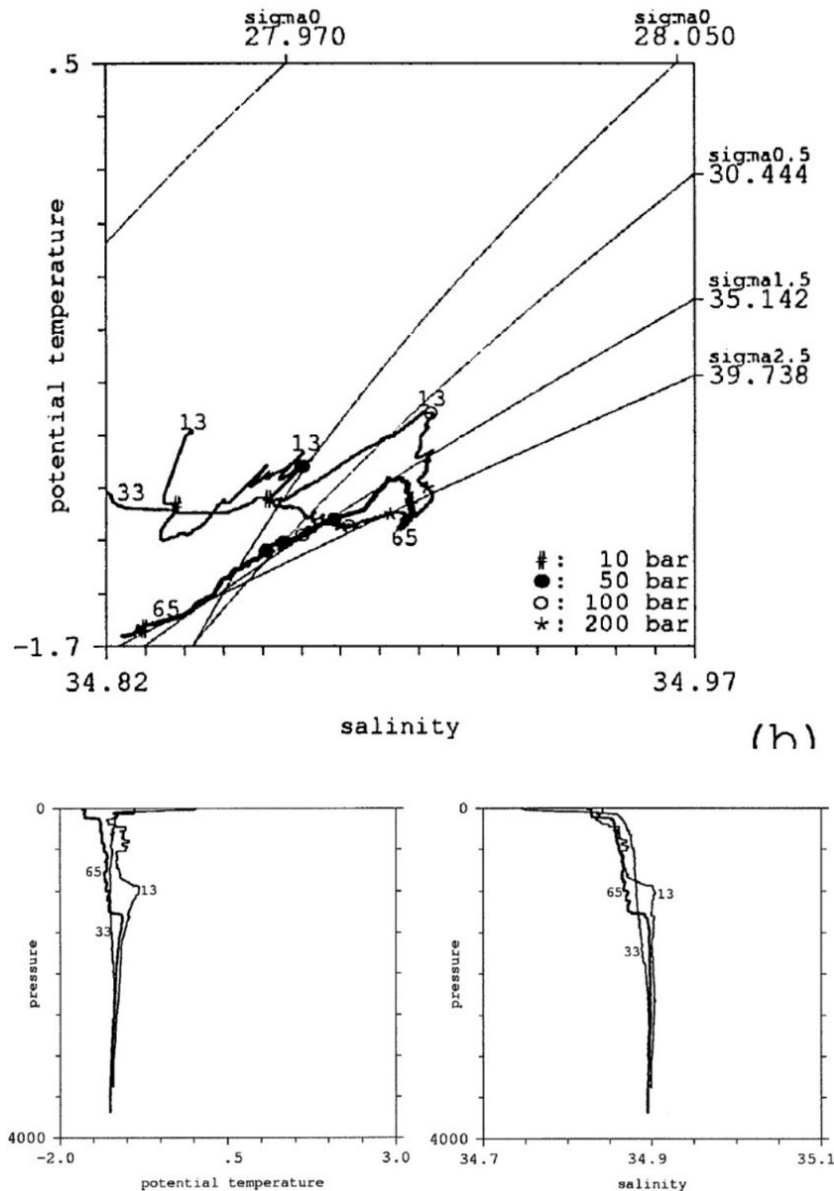


Figure 14. Potential temperature and salinity profiles and θ - S curves obtained by FS *Valdivia* in 1988 (65 and 33) and 1993 (13) in the central Greenland Sea, showing the evolution of the convection during the winter of 1988 and the shallower convection observed in 1993 (adapted from Rudels et al., 1999b).

deep water to the center of the gyre and the vertical mixing into the bottom water. The obtained mixing coefficient was surprisingly high, $\kappa_v \approx 3.7 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$. The renewal time for the deep salinity maximum was estimated to 6 years, corresponding to a through flow of 0.3 Sv. If the removal of the Greenland Sea bottom water, 0.2 Sv, is included 0.5 Sv of deep water is exported from the Greenland Sea (Meincke et al., 1997). Later studies made by Somavilla et al. (2013) and Marnela et al. (2016) have indicated similar high vertical mixing coefficients.

The changes in the Greenland Sea water column continued during the 1990s and a temperature maximum started to develop at mid depth (Budéus et al., 1998), indicating that the Amerasian Basin deep water penetrating towards the center of the Greenland Sea was no longer removed or weakened by mixing with convecting cold, less saline Arctic intermediate water. The water locally formed in winter has become less dense and only the water column above the temperature maximum is ventilated by convection. The temperature maximum, by contrast, is renewed by advection, locally from the Greenland slope but ultimately from the Arctic Ocean, which

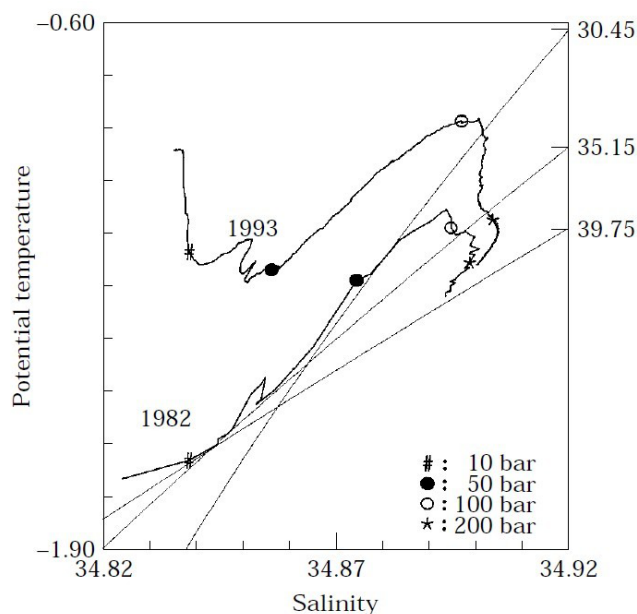


Figure 15. θS curves from two stations taken 11 years apart, by RV *Hudson*, 1982 at 75°02' N, 005°11' W and by FS *Valdivia*, 1993 at 74°00' N, 004°20' W (Figure from Meincke et al., 1997).

gradually increases its temperature. The temperature maximum also moved downwards in the water column (Budéus and Ronski, 2009), implying that the local convection, although less dense than the water of the temperature maximum, did not stop sinking when it reached its neutral density level. Instead, its layer thickness increased, indicating that it was pressing the underlying water column downwards, by continuity forcing the bottom water towards the rim and eventually to leave the Greenland Sea.

This radically changed the density structure of the central Greenland Sea. Instead of displaying a dome of dense water reaching towards the surface the density surfaces formed a bowl depressed in the center and rising towards the rim. This change implies that the doming, which was earlier considered generated by wind forcing and a necessary preconditioning for convection (Killworth, 1983; Marshall and Schott, 1999) instead is a consequence of open ocean convection. When deep convection ceases, the doming cannot be maintained and the density surfaces relax.

During this phase deep reaching cold, vertically almost homogenous anticyclonic vortices were observed in the Greenland Sea (Gascard et al., 2002; Wadhams et al., 2002). Their lateral extent was 10 to 15 km, close to the local Rossby radius, which prevented them from breaking up. The vortices reached deep (> 2000 m) and penetrated into the temperature maximum. The upper part of the vortices was denser but their lower part less dense than the ambient water column. The vortices were, as the rest of the Greenland Sea, capped by a low salinity layer in summer. However, the large ver-

tical extent of the vortices made this upper layer less thick and it would more rapidly become dense in winter and convect into the vortices, recharging them with dense water. This early started and confined convection could also be a cause for the deep vertical penetration of the vortices. The fact that the vortices were less dense than the ambient water between 1500 and 2000 m made them, however, unlikely candidates for a renewal of the Greenland Sea deep and bottom water, but they stand out as intriguing features in the water column. Vortices were observed during a 10 years period but then disappeared.

In 1996 a tracer release experiment to study the convection was conducted in the central Greenland Sea. A cloud of Sulphur Hexachlorid was injected in the central gyre at the density surface $\sigma_{0.5} = 30.4268$ and its spreading and mixing were monitored during the following years (Watson et al., 1999). The first winter was one with comparably strong ice formation, which likely influenced the mixing of the tracer. Outside the gyre the tracer cloud spread but little vertically but in the gyre the tracer was redistributed vertically from the surface down to 1300 m. The highest tracer concentration was reduced and the maximum was shifted slightly upward.

One interpretation of these changes is that cooling or ice formation and brine rejection create denser water at the surface that convects into the deep as entraining plumes that pass through the tracer cloud and redistribute the tracer to deeper levels. By mass continuity a compensating upward flow through the tracer cloud then brings the tracer also to shallower levels. The one-dimensional convection model applied by Watson et al. (1999) agrees with this picture.

With time the temperature of the cold Arctic intermediate water above the temperature maximum gradually increased and eventually it became higher than the temperature of Amerasian Basin water, which instead only was identified as a sharp salinity increase below the ventilated intermediate water. By mid 2010s the ventilated intermediate water had also become more saline than the Amerasian Basin water and its salinity maximum disappeared.

This is a likely consequence of the increase in salinity and temperature of the upper layer in the Greenland Sea due to interaction with recirculating Atlantic water in the East Greenland Current, which counteracts the atmospheric cooling taking place in the central gyre (Latarius and Quadfasel, 2016). The salinity and temperature of the upper layer in the Greenland Sea are now the highest observed in the Greenland Sea water column. This has made the Greenland Sea similar to the Labrador Sea and the convection now resembles that of a gradual cooling and deepening of the upper layer in winter.

The cold, less saline Greenland Sea bottom water has also disappeared and the θS characteristics of the bottom water are now those of the Eurasian Basin deep water (Marnela et al., 2016). The features of the water column that identified two competing sources of the deep water in the Arctic Mediterranean (Aagaard et al., 1985; Rudels, 1986a) have disappeared, and presently the Arctic Ocean source domi-

nates and the Greenland Sea deep water is renewed by advection. If the Arctic Ocean source also has weakened, it has not yet been detected in the Greenland Sea because the larger reservoir present in the Arctic Ocean. The cause behind these changes is likely the warming climate that no longer creates water dense enough to ventilate the deep Greenland Sea by open ocean convection.

11 Overflows

The Arctic Ocean and Greenland Sea deep waters are mostly confined to the Greenland Sea by the Jan Mayen fracture zone and the Mohn–Knipovich Ridge and contribute but little to the overflows. When the intensive studies of the Nordic Seas started in the 1970s and 1980s an outflow of Greenland Sea deep water to the Norwegian Sea was observed in the about 2200 m deep channel in the Mohn Ridge just north of Jan Mayen. In the late 1990s this flow had changed direction and instead deep water from the Norwegian Sea entered the Greenland Sea, implying that the Greenland Sea deep water density had been reduced and that the deep pressure gradient had reversed (Østerhus and Gammelsrød, 1999). Since no deep water is formed in the Norwegian Sea, this indicates that water, earlier exported from the Greenland Sea, was returning. This situation probably persisted only a short period since, due to continuity, the deep water from the Arctic Ocean entering the Greenland Sea has to leave, either through gaps in the Mohn–Knipovich Ridge to the Norwegian Sea or return to the Arctic Ocean through Fram Strait.

The deepest passage across the Greenland–Scotland Ridge to the North Atlantic is the Faroe Bank Channel (> 840 m) and the densest deep water that enters the Norwegian Sea from the Greenland Sea must eventually exit the Arctic Mediterranean through this passage. For less dense water there is also the possibility of spillover across the 500 m deep ridge segment between Iceland and the Faroes. The Faroe Bank Channel is a continuation of the Faroe–Shetland Channel and Fig. 16, showing bottom temperature and bottom salinities in the northeastern North Atlantic, reveals the narrowness of the channel and also the spillover of cold, less saline water across the Iceland–Scotland Ridge. Not all dense water that passes through the Faroe–Shetland Channel reaches the Faroe Bank Channel. Some less dense water continues south and crosses the Wyville Thomson Ridge and enters the Rockall Basin. As the name of the ridge suggests, this is where Carpenter and Wyville Thomson first observed the cold water entering the Northeast Atlantic from the north.

Not all the overflow water from the Norwegian Sea derives from the Greenland Sea. A substantial fraction of the East Greenland Current that crosses the Jan Mayen Fracture Zone and enters the Iceland Sea continues eastward in the East Icelandic Current into the Norwegian Sea. The East Icelandic Current is deep, about 600 m, and carries not only low salinity surface water but also denser waters that especially

contribute to the spillover across the Iceland–Faroe Ridge. In a drifter experiment with RAFOS floats in the Iceland Sea with a drifting depth of 600 m all of the floats deployed east of the Kolbeinsey Ridge moved southeastward into the Norwegian Sea, and also some deployed west of Kolbeinsey Ridge entered the Norwegian Sea. However, most of the floats deployed west of the ridge either exited through Denmark Strait or grounded on the northwestern Icelandic shelf (De Jong et al., 2017).

The eastern overflow entrains strongly the ambient waters and as the overflow reaches its neutral density level the overflow volume has more than doubled. The overflow water moves toward the equator but when it reaches the Gibbs Fracture Zone in the Mid-Atlantic Ridge it splits and some, maybe the major part, leaves the Iceland Basin and enters the Irminger Basin and then moves north towards Iceland. The initial low temperature and salinity characteristics of the eastern overflow has rapidly been removed and it has attained a potential temperature between 2 and 3 °C and a salinity around 34.925.

The western overflow in Denmark Strait is more complicated. It is not a deep, almost stagnant reservoir that gradually drains south over a sill, but instead a region where several current branches and water masses converge. The upper part of the East Greenland Current that passes over the sill depth of the Jan Mayen Fracture Zone (1700 m) comprises not only low salinity Polar water, but also Arctic Atlantic and denser intermediate waters from the Arctic Ocean, as well as warm and saline Atlantic water recirculating in Fram Strait and Arctic Intermediate waters created in and exported from the Greenland Sea. There is also the recently discovered North Icelandic Jet roughly following the 600 m isobath on the northern Iceland slope westward to Denmark Strait (Jónsson and Valdimarsson, 2004). In addition, the warm and saline North Icelandic Irminger Current enters the Denmark Strait from the south. The sill depth in Denmark Strait is 640 m, but the deep channel is narrow and waters from north and south contend for space and passage. Changes take place from one week to the next and Fig. 17 shows two sections along the sill taken by RV *Aranda* one week apart. On the first section waters from the north dominates, on the second water from the south. The second section extended farther towards Greenland because of favorable ice conditions (Rudels et al., 1999c).

Overflow water is commonly defined as water with potential density above 27.8 and the second section shows that waters with density above 27.8 are present on the shelf, which implies that even when the North Iceland Irminger Current dominates in the deep channel the overflow, or at least part of it, could instead take place over the shelf and then sink down the continental slope into the Irminger Basin. However, the East Greenland Current is not the only source of overflow water. In fact, the densest water crossing the sill is supplied by the North Icelandic Jet. The jet has been traced eastward

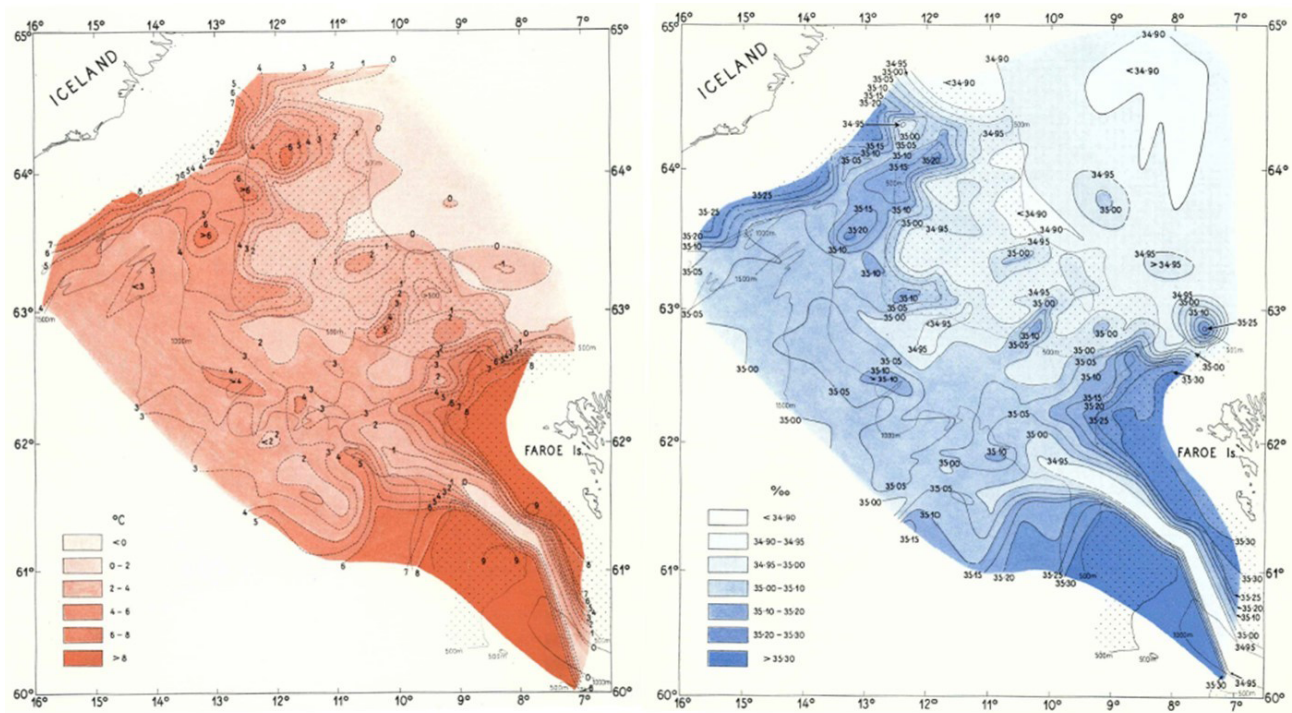


Figure 16. The bottom temperature (left) and salinity (right) distributions observed between Iceland and the Faroes during ICES's Overflow-60 experiment. Especially the cold and less saline water flowing through the Faroe Bank Channel is conspicuous compared to the cold water spilling over the Iceland Faroe Ridge and clearly reveal the narrowness of the channel (from Tait et al., 1967).

towards its unknown source area and it is observed along the entire northern slope (Våge et al., 2011).

Våge et al. (2011) suggested that the water of the North Icelandic Jet is supplied by two sources, the first would be the cooling of the North Icelandic Irminger Current as it flows eastward on the shelf. The second contribution would come from mixing with the denser water in the interior of the Iceland Basin as the cooled water returns towards Denmark Strait, further increasing its density. The problem with this explanation is that the atmospheric cooling observed in the area is not strong enough to create the observed θS properties of the jet (Segtnan et al., 2011; Latarius and Quadfasel, 2016). Furthermore, the seasonal convection in the Iceland Sea is usually limited to the upper 200 m and only in the northern part of the Iceland Sea does the winter mixed layer occasionally attain the θS observed in the North Icelandic Jet (Våge et al., 2015).

That the formation of dense water in the Iceland Sea in winter and its subsequent export in summer and fall was the main process supplying the Denmark Strait overflow was proposed by Swift et al. (1980) and Swift and Aagaard (1981) as an alternative to the cooling of the Atlantic water in the Norwegian Sea and its recirculation in Fram Strait and export in the East Greenland Current suggested by Worthington (1970). However, Mauritzen (1996a, b) did not find any strong seasonality in the overflow, which speaks against

seasonal sources in the central basins of the Nordic Seas and rather indicates a steady flow, which could be supplied by the gradually cooled Atlantic water returning to the North Atlantic through Denmark Strait.

Waters with the θS characteristics of the deepest overflow are present in the northern Iceland Sea and in the East Greenland Current. These θS properties are, however, observed at much deeper levels, 750 to 1500 m, than the about 500 m depth, where the core of the North Icelandic Jet is found. Moreover, it is likely that these waters are formed farther north and have been advected across the Jan Mayen Fracture Zone to the Iceland Sea. If the East Iceland Current transports a major part of the upper 600 m of the water column to the Norwegian Sea, the deeper levels might, if they are advected southward, ascend to shallower depth as they approach the Iceland Continental slope and then feed the North Icelandic Jet (Fig. 18).

As a curiosity, the deep salinity maxima shown in Fig. 18 derive from the Jan Mayen Fracture Zone (blue station) and the central Greenland Sea (red station) and have about the same density, but the maximum at the station at the rim is located at shallower depth than the maximum at the station in the central Greenland Sea. This implies that the convection in the Greenland Sea has compressed the underlying water column, creating the deep part of a bowl, which then rises towards the rim. These stations were taken in 2002, when

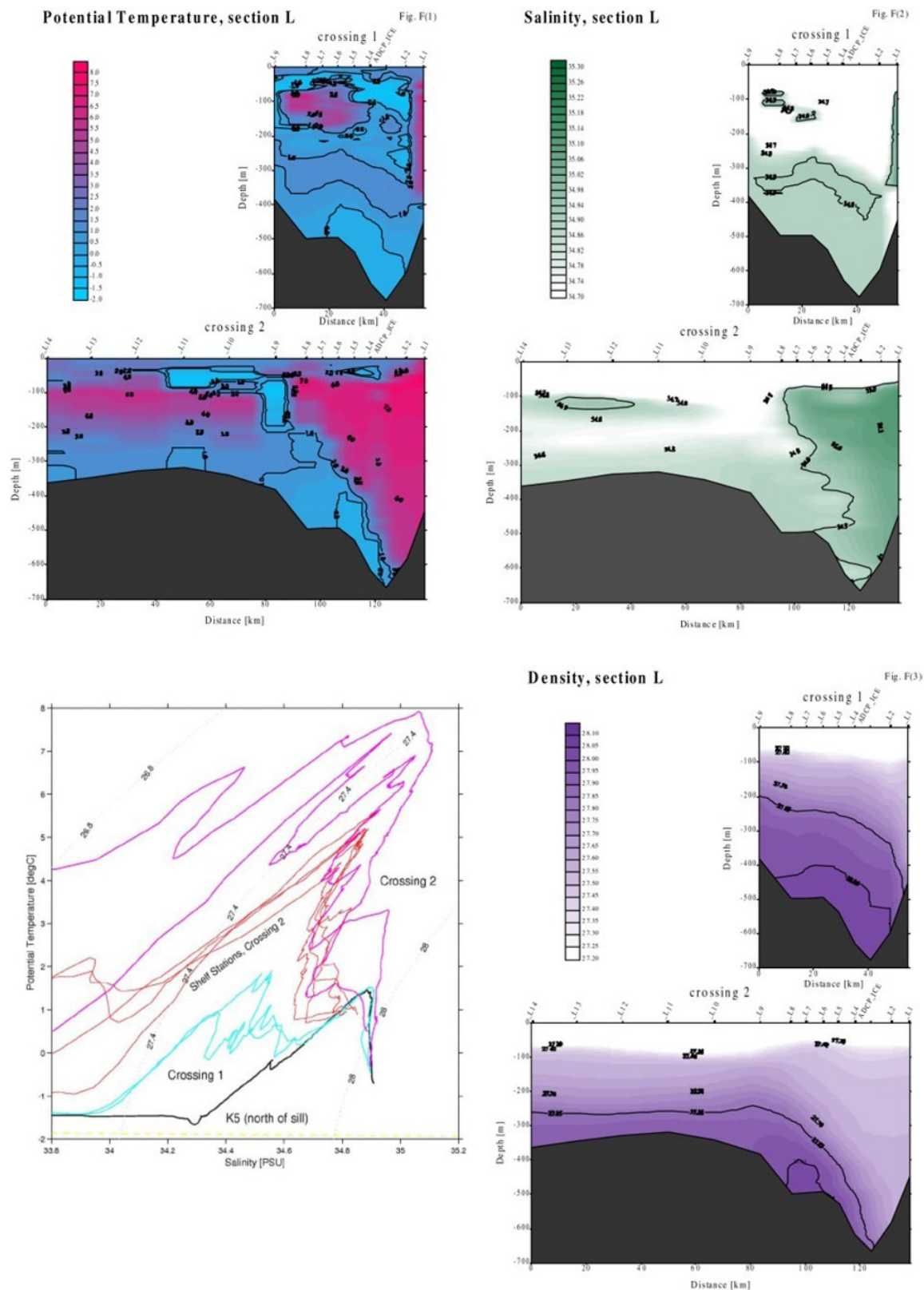


Figure 17. Potential temperature, salinity and potential density sections from two crossings between Greenland and Iceland along the sill in Denmark Strait taken in 1997 one week apart. The θS diagram shows the θS curves from the two crossings, the first crossing in cyan and the second crossing in red and magenta. (The figure is adapted from Rudels et al., 1999c.)

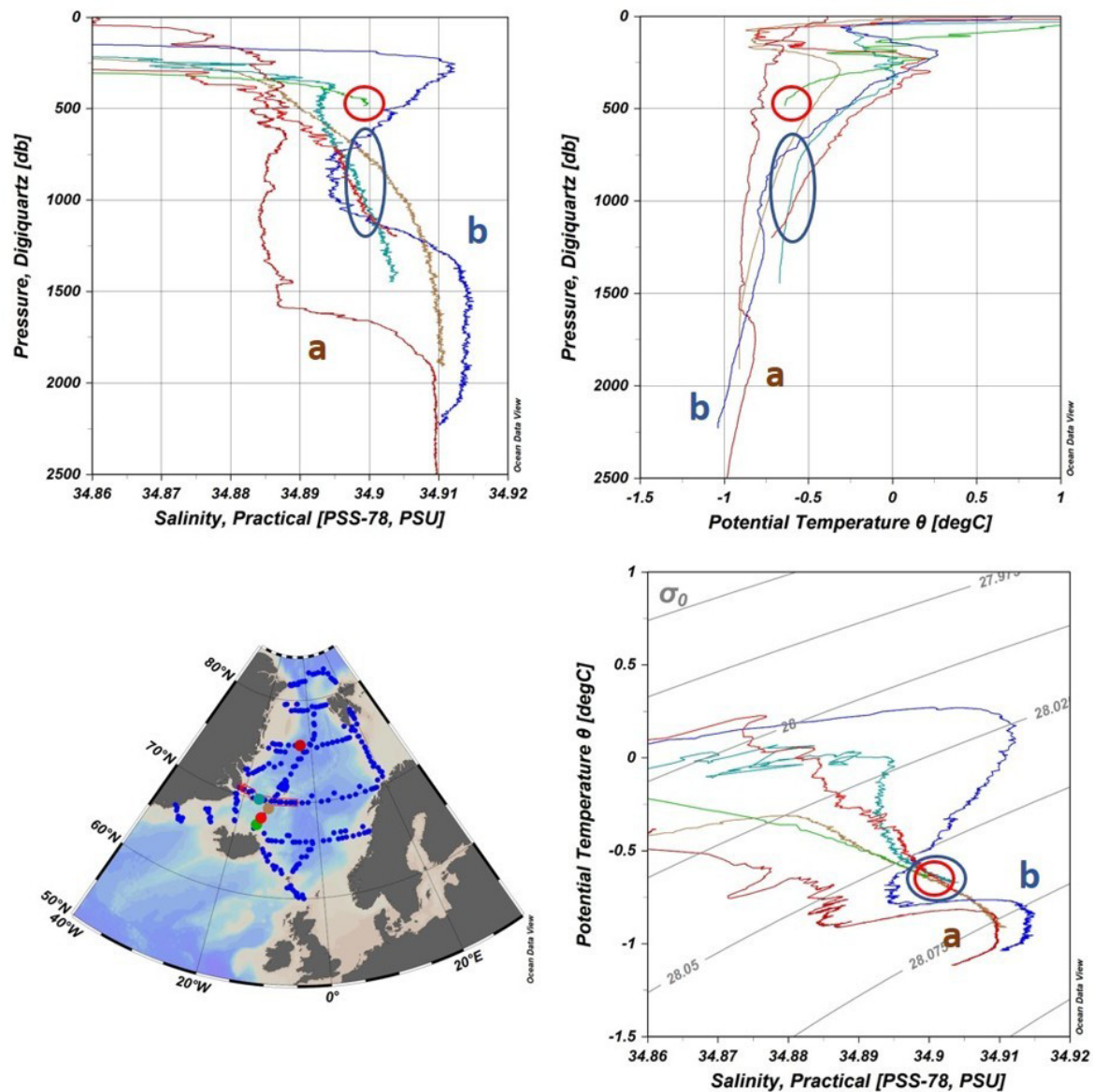


Figure 18. Profiles of potential temperature and salinity and θS curves from stations in the Iceland Sea and the Greenland Sea. The green station (red circles) found at 480 m depth at the Iceland slope has similar θS properties as the densest overflow and also properties akin to those observed below 750 m and still deeper on stations farther north in the Iceland Sea, indicated by blue ellipses. These properties are located at deeper levels, if a warmer Atlantic layer also is present, suggesting that it is pushed downward by the Atlantic water. The θS curves from these stations have similar shape as the upper Polar deep water (temperature decreasing and salinity increasing with depth), but are colder and located lower in the θS diagram. The θS properties could be created by mixing between the upper Polar deep water and the less saline and colder Arctic Intermediate water formed in the Greenland Sea. However, the smooth θS curves do not indicate recent interleaving and mixing but rather suggest long residence time and slow renewal. Stations a and b shows the difference between the central gyre and the southern rim of the Greenland Sea. The observations are from RV *Knorr* 2002 (Figure from Rudels, 2021).

the Greenland Sea convection had become shallower, and it should be compared with the stations in Fig. 13 in Sect. 10, taken in 1988, when the denser water is found at shallower levels in the center than at the rim.

The overflow plume in Denmark Strait is stratified. It comprises a cold, $< -0.5^{\circ}\text{C}$, bottom layer with salinity

≈ 34.9 below a temperature and a salinity maximum of 2°C and 34.91, respectively, and it is capped by a salinity minimum, a low salinity lid (Fig. 19). As the plume leaves the sill its bottom temperature increases rapidly to above 0°C but remains below 1°C even when the plume has descended deeper than 3000 m in the Irminger Sea, and the low salin-

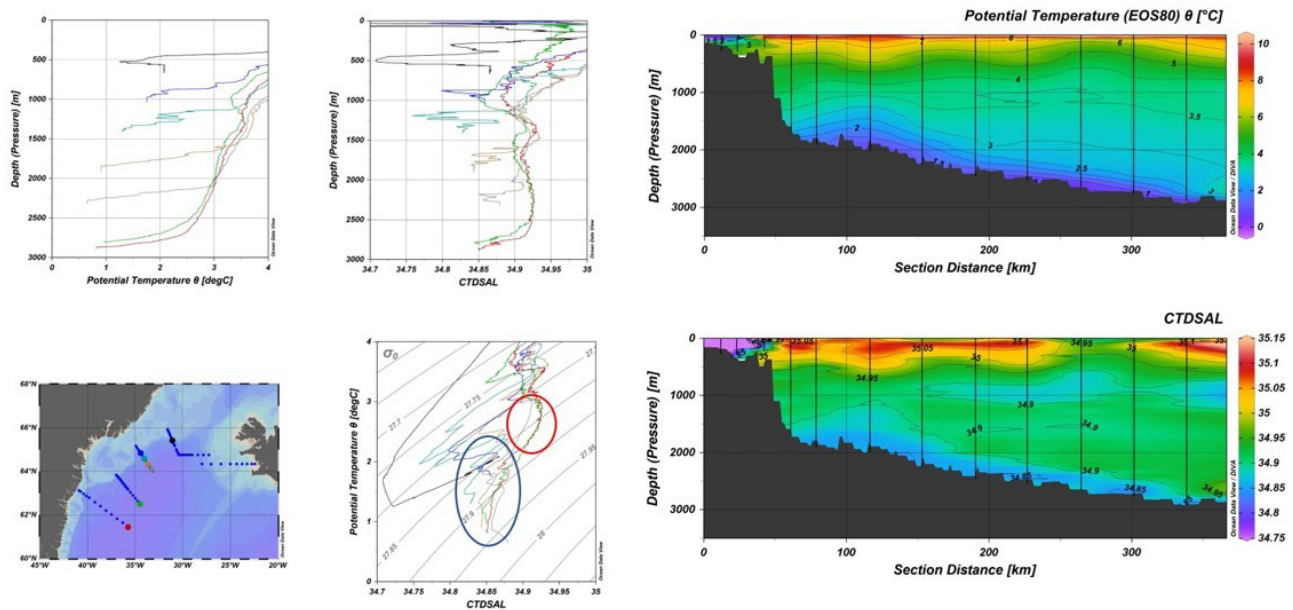


Figure 19. Potential temperature and salinity profiles and θS curves from the Greenland slope south of Denmark Strait. The southernmost section shows the overflow as a cold, less saline bottom layer, the warm Irminger Current in the upper layer and the presence of the more saline and warmer Iceland-Scotland overflow water below 2000 m. The profiles and the θS curves reveal the downstream evolution of the overflow properties, especially the low salinity lid but also the difference between the two overflows, the Iceland-Scotland overflow indicated by red ellipse and the Denmark Strait overflow by blue ellipse in the θS diagram. The potential temperature and salinity distributions on the southernmost section are shown (data from RV *Alexander von Humboldt* 2004) (Figure from Rudels, 2021).

ity lid is present almost as deep. The initial rapid increase in bottom layer temperature suggests that a direct mixing between the overflow water and the northward flowing North Iceland Irminger Current could take place already at the sill in Denmark Strait. There the deepest and densest part of the overflow comes in almost direct contact with the warm and saline Irminger Current water, and the observed increase in plume temperature in the deep Irminger Basin could result from a mixing ratio at the sill of one part of entrained water to four parts of overflow water. This is significantly less than if the entrainment would occur deeper down the slope. It is also less than the estimated entrainment for the Faroe Bank overflow.

The temperature minimum and the low salinity lid, by contrast, must derive from the Greenland shelf, where waters of the observed properties are found. The fact that especially the low salinity lid survives as the plume descends deeper than 2000 m also indicates that the entrainment of ambient water through the upper boundary of the plume is weak. However, the low salinity lid suggests that if the less dense water can remain in contact with the densest part of the plume in the early stages, when it starts to descend the slope, it will remain attached to the plume and sink into the deep. One factor that can facilitate the descent of the cold low salinity upper part of the plume is the thermobaric effect, which increases the relative density of colder water as the pressure increases. As the Denmark Strait plume descends it eventually sinks

through the overflow water from the Faroe Bank channel that shows up as a salinity maximum above the colder, less saline Denmark Strait overflow. Both overflows then continue to the Labrador Sea.

If the dense water that flows southward on the Greenland shelf reaches the shelf break farther south, where the shelf narrows, the entrainment appears to be stronger and the plumes will remain on the upper part of the slope and not connect to the deeper overflow. This could be the case for the Greenland spill jet, identified by the density surfaces rising towards the shelf, which indicates velocities increasing with depth (Pickart et al., 2005). It flows as a separate core above the deep overflow plume towards Cape Farewell and eventually the density surfaces flatten and the spill jet disappears (Pickart et al., 2005).

The overflow in Denmark Strait is strong and varies on time scales of a few days (2 to 5 d). The strong overflow events were noticed by Worthington (1969) as the first direct current meter array was recovered. These strong overflow events had periods slightly less than a week. However, later observations (Von Appen et al., 2017) have shown that two different kinds of strong overflow events are present. The boluses are related to overflow events where the thickness of the overflow plume increases drastically, the temperature of the boluses decreases, but no significant increase in the overflow velocity is observed. The second overflow anomaly Von Appen et al. (2017) denoted as pulses. The pulses show an in-

crease the overflow velocity but the thickness of the overflow layer decreases. The temperature reduction is less (-0.1°C) than that observed for the boluses (-0.4°C) and the more rapid flow in the pulses compensates for the thinner overflow plume and the overflow strength is about the same for the boluses and the pulses.

The difference of height of the overflow plumes first suggests a supercritical flow for the boluses and a subcritical flow for the pulses, but nature is most likely more devious than this. The flow is rotating and the thickening of the boluses is associated with cyclonic rotation and the thinning of the pulses relates to anticyclonic rotation (Von Appen et al., 2017). The most interesting question is, however, if the boluses and pulses are associated with different water masses. The almost similar temperature of the overflow water in the two cases suggests that the main sources are in the central gyres, not in the East Greenland Current, and rather in the Greenland Sea than in the Iceland Sea.

The overflows contribute together with the dense water formed by open ocean convection in the Labrador Sea to the North Atlantic deep water and to the Atlantic Overturning Circulation. The overturning circulation has been monitored since 2004 by the RAPID-MOCHA (Meridional Overturning Circulation and Heat-flux Array) array deployed at 26°N . One of the drawbacks with this array is that it is located too far south to separate the overturning circulation, sinking into the deep, from the horizontal recirculation of cooled upper water in the subtropical gyre. An additional array, OSNAP (Overturning in the Subpolar North Atlantic Program), located north of the subtropical gyre has therefore been proposed and accepted, and the first deployment was in 2014. The array extends from Scotland to Greenland (OSNAP east) and from Greenland to Labrador (OSNAP west). The first deployment ended in 2016 and when the results from the first 20 months were published (Lozier et al., 2019) the title “A sea change in our view of overturning in the subpolar North Atlantic” suggests that the results, at least for some of the researchers involved, have come as a surprise. The convection and dense water formation in the Labrador Sea contribute, as estimated by the OSNAP west array, but little, $2.1 \pm 0.3 \text{ Sv}$, to the overturning circulation. By contrast the transformation to denser water measured by the OSNAP east array was $15.6 \pm 0.8 \text{ Sv}$, almost an order of magnitude larger. Since some recirculation takes place, the measured total overturning is somewhat less, $14.9 \pm 0.9 \text{ Sv}$.

For those who have been working in the Arctic Mediterranean Sea and on the exchanges across the Greenland–Scotland Ridge these numbers are not surprising and should also be familiar to those remembering L.V. Worthington’s (1976) monograph on the North Atlantic circulation. The observations made during the NACLIM project (North Atlantic CLIMate) indicated an overflow of 5.8 Sv (Østerhus et al., 2019). Assuming a doubling of the volume due to entrainment into the plumes and some contribution from the spill jet the numbers almost match.

The small volume of Atlantic deep water formed in the Labrador Sea comes as a surprise though. It suggests that seasonal deep convection is not a process that creates and drives an overturning circulation. It forms MODE waters rather than deep water (McCartney, 1977; McCartney and Talley, 1982). To create the dense water that supplies the overturning circulation, a sill and an isolated basin might be required. This would inject water with substantially higher density contrast that can sink down and entrain ambient water, creating a significant volume of deep water with distinct properties. The fact that also the much weaker Mediterranean overflow shows up in and characterizes the Atlantic water column over a large region suggests that this could be the case.

Observations from the RAPID array indicates a reduction of the overturning circulation by 2.5 Sv in 2009. The weaker overturning has persisted but no further reduction has been observed (Bryden et al., 2020). The reduction has, surprisingly, taken place in the denser part of the overturning circulation, which is supplied by the overflow from the Arctic Mediterranean. During the same time the EU projects THOR (ThermoHaline Overturning at Risk) and NACLIM were running and no reduction was observed, not in the overflows, nor in the northward inflows of Atlantic water to the Arctic Mediterranean.

The RAPID array showed that reduction of the overturning circulation was compensated by an increase in the recirculation in the subtropical gyre by 2.5 Sv , about 15 % of the overturning circulation. This implies that less warm Atlantic water is transported by the North Atlantic drift to the subpolar gyre, which should lead to a cooling of the northeast Atlantic.

12 Polar outflows and the Canadian Arctic Archipelago and Baffin Bay

The water leaving the Arctic Mediterranean Sea consists not only of dense overflow water but also of cold, less saline and less dense polar water that interacts with the upper part of the subpolar North Atlantic. There are two main, interacting, outflows, the outflow of Polar Water through Fram Strait in the East Greenland Current that follows the Greenland shelf and coast southward through Denmark Strait to Cape Farewell, and the polar water that flows through the narrow and shallow passages in the Canadian Arctic Archipelago to Baffin Bay and then continues through Davis Strait to the Labrador Sea. These two flows interact. A part of the East Greenland Current turns north after it has rounded Cape Farewell and forms the West Greenland Current that enters Baffin Bay from the south. The upper, polar, part of West Greenland Current mixes with the outflows from Nares Strait, Jones Sound and Lancaster Sound and the mixing product is transformed by ice formation during winter into a distinct, about 200 m deep upper layer with temperature of -1.6°C and salinity of 33.7 (Fig. 20).

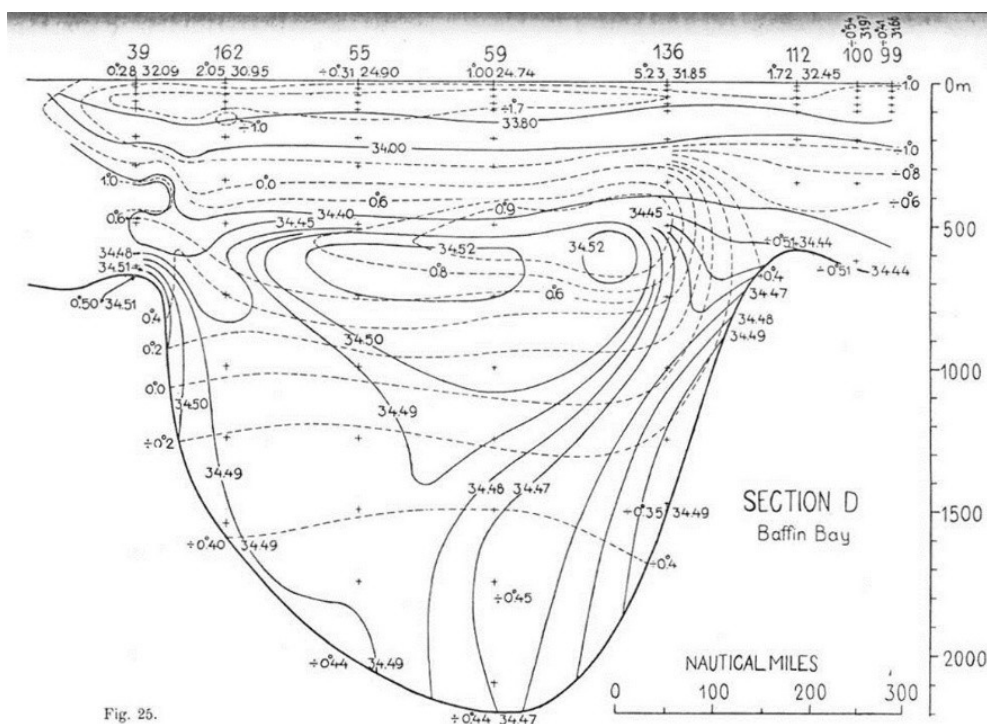


Fig. 25.

Figure 20. Temperature (dotted lines) and salinity (full lines) sections taken by *Godthaab* 1928 in Baffin Bay from Davis Strait (left) to Smith Sound (right), showing the cold upper layer, the intermediate temperature and salinity maxima and the cold, slightly less saline bottom water (from Riis-Carstensen, 1936).

Below the low salinity upper water, a warm, 3–4 °C, but rather fresh, 34.5–34.6, Atlantic layer is present. The Atlantic water enters Baffin Bay through Davis Strait in the West Greenland Current and the low salinity indicates that it was originally a part of the Irminger Current that has become diluted by cold, low salinity water from the East Greenland Current south of Denmark Strait.

The bottom water in Baffin Bay is again colder, about −0.45 °C with salinity 34.45. The origin of the Baffin Bay bottom water has been debated, Sverdrup et al. (1942) considered it to be a mixture between Labrador Sea water and Baffin Bay surface water, and local formation by convection was suggested by Bourke and Paquette (1991). Bailey (1956) found TS characteristics at sill depth north of Nares Strait similar to those of the Baffin Bay bottom water and proposed that bottom water entered from the Arctic Ocean. Figure 21 shows the TS curves from one station in Baffin Bay and two stations from the Arctic Ocean, one from the continental slope in the Canada Basin and the other from the slope northeast of Greenland, indicating that the Baffin Bay bottom water could derive from the Arctic Ocean.

The TS curves also show the warm but less saline characteristics of the Baffin Bay Atlantic water compared to the Arctic Atlantic water to the north. They also, interestingly, indicate that the entrainment of ambient water into the deep inflow must be small. If the inflow entrains much, it will be-

come warmer, which reduces its density. This, in turn, requires a denser initial inflow. However, that would also mean a warmer inflow, which would lead to different TS characteristics in the Baffin Bay bottom water.

To determine the export of low salinity polar water by direct current measurements is difficult. Drifting ice and bad weather conditions complicate not only the deployment and recovery but also the survival of the moorings. Since one primary object is to determine the freshwater fluxes that may influence the convection in the subpolar gyre, it is essential to obtain measurements close to the surface and the ice, where most of freshwater is located. In spite of these difficulties measurements of volume and freshwater fluxes have been made, at least for shorter periods, in all the major passages; Lancaster Sound (Peterson et al., 2012), Jones Sound (Cardigan Strait and Hells Gate) (Melling et al., 2008), and Nares Strait (Münchow and Melling, 2008; Rabe et al., 2010) in the Canadian Arctic Archipelago, in Davis Strait (Curry et al., 2011, 2014), and on the shelf area east of Greenland (De Steur et al., 2009; De Steur et al., 2016). The techniques used in the different passages are distinct and adapted to the local challenges such as the presence of sea ice and the closeness to the magnetic North Pole that complicates the direction measurements. Details of these observations are given in Melling et al. (2008) and in Rudels (2021).

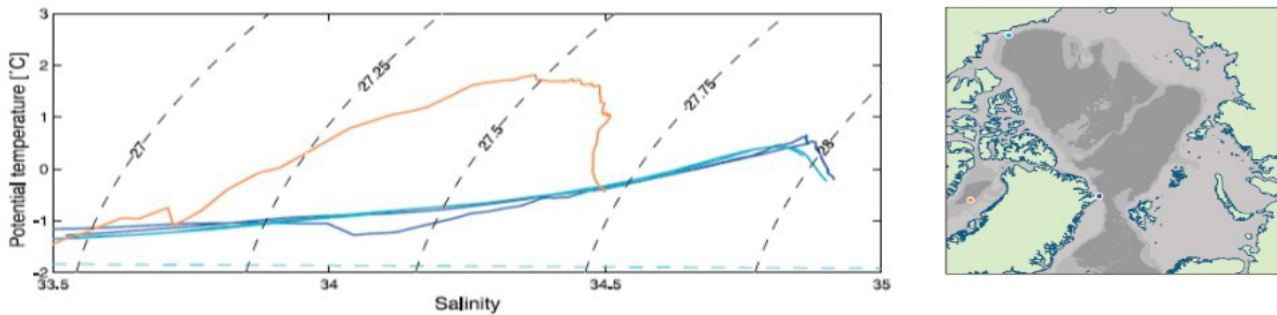


Figure 21. TS curves from Baffin Bay (yellow) and from north of Canada (cyan) and north of Greenland (blue), showing that the Baffin Bay bottom water could derive from the Arctic Ocean (from Rudels et al., 2004).

The difficulty to perform field observation, especially for monitoring purpose, makes it tempting to use existing simple models to estimate the transports of Polar water through the different passages. One approach follows Werenskiöld (1935), who derived an expression for the geostrophic transport M in a coastal current based only on the depth of the upper layer at the coast, H_1 , and the density difference ($\rho_2 - \rho_1$) between the upper layer and the lower layer at rest, which extends to the surface one Rossby Radius away from the coast.

$$M = \frac{(\rho_2 - \rho_1)gH_1^2}{2f\rho_2} \quad (4)$$

The density in cold Arctic waters is mainly controlled by salinity and can be written as $\rho = \rho_r(1 + \beta S)$, where ρ_r is the reference density and β the coefficient of salt contraction. Taking the density of the lower layer as the reference density the reduced gravity ρ' becomes $g\beta(S_2 - S_1)$.

This represents the upper circulation and the amount of lower layer Atlantic water that enters the Arctic Ocean and is entrained into the upper layer. Stigebrandt (1981) modelled the entrainment with a Kato–Phillips mixing model and noticed that if the openings are wider than a Rossby radius they all transport with the same rate.

There are three straits that are wider than the Rossby Radius and can transport at full capacity, Fram Strait, Lancaster Sound and Nares Strait. The total transport through these passages comprises the liquid freshwater flux F added to the water column by rivers and by the low salinity Pacific inflow, and the entrained lower layer water M_2 . The Pacific inflow is here separated into freshwater and lower layer water fluxes. This gives:

$$M_2 + F = \frac{3g\beta(S_2 - S_1)H_1^2}{2f} \quad (5)$$

and the thickness of the freshwater m in the upper layer becomes:

$$m = \frac{(S_2 - S_1)}{S_2} H_1 \quad (6)$$

The salinity of the upper layer can be written as:

$$S_1 = \frac{M_2 S_2}{(M_2 + F)} \quad (7)$$

And introducing this in the expression for the outflow gives:

$$(M_2 + F) = \left(\frac{3g\beta F S_2}{2f} \right)^{\frac{1}{2}} H_1 \quad (8)$$

and m becomes:

$$m = \frac{S_2(M_2 + F) - (M_2 S_2)}{S_2(M_2 + F)} H_1 = \frac{F H_1}{(M_2 + F)} = \left(\frac{2fF}{3g\beta S_2} \right)^{\frac{1}{2}} \quad (9)$$

The Rossby Radius Ro is independent of the entrainment and only depends on the freshwater input:

$$Ro = \frac{[g\beta(S_2 - S_1)H_1]^{\frac{1}{2}}}{f} = \left(\frac{2g\beta S_2 F}{3f^3} \right)^{\frac{1}{4}} \quad (10)$$

The Rossby radius is about 10 km, which confirms that both Lancaster Sound and Nares Strait are wide enough to transport with full capacity.

If we assume that the depth H_1 is the same in all channels and equal to the depth of the upper layer in the Arctic Ocean, the liquid freshwater storage in the Arctic Ocean can be estimated. Introducing $g = 9.83 \text{ m s}^{-2}$, $f = 1.4 \times 10^{-4} \text{ s}^{-1}$, $\beta = 8 \times 10^{-4}$, $F = 0.20 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ and setting S_2 equal to 35 and the salinity of sea ice to zero the liquid freshwater storage becomes 8.24 m. The area of the Arctic Ocean is about $10 \times 10^{12} \text{ m}^2$ and with a freshwater export of 0.2 Sv the residence time becomes 13 years, or half of this if only the deep part of the Arctic Ocean is considered. The freshwater storage increases as $F^{1/2}$ but its residence time decreases as $F^{-1/2}$ (Rudels, 2010).

The thickness of the ice cover was in the 1960s estimated to 3 m, about one third of the liquid freshwater storage but had decreased by about 40 % by the 1990s (Rothrock

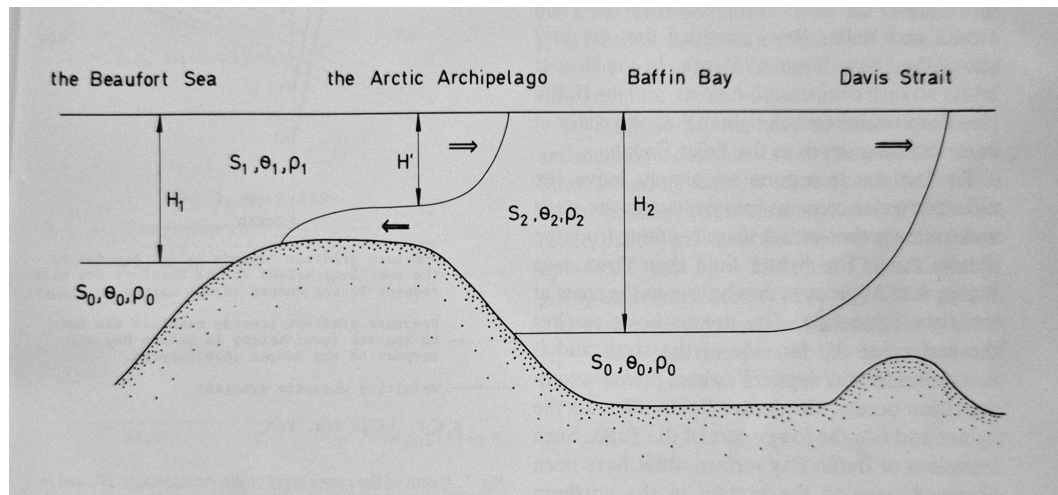


Figure 22. Idealized three reservoir system representing the Beaufort Sea, the Canadian Arctic Archipelago, Baffin Bay, Davis Strait and Labrador Sea system. S is salinity, θ potential temperature and ρ potential density. Subscript 1 indicates polar surface water, subscript 2 Baffin Bay upper water and subscript 0 indicates the deep water in the Arctic Ocean and in Baffin Bay that communicates freely through Fram Strait, across the Greenland–Scotland Ridge and over the sill in Davis Strait (from Rudels, 1986b).

et al., 1999). If the ice export $F_i = 0.09$ Sv, remains constant, the mean residence time of the ice cover is between 2.6 and 5.3 years depending upon the area considered. If only a seasonal ice cover were present and no sea ice were exported the liquid freshwater storage would increase to a little above 10 m (Rudels, 2010). The liquid freshwater in the Arctic Ocean is, however, not a lid lying on top of a deeper water mass but is distributed over a large, spatially varying, depth range in the Arctic Ocean water column. The conditions in the three passages are also not equal. Especially the characteristics of the second layer differ between the passages, which affects the transports of both volume and freshwater.

The second passageway from the Arctic Ocean to the North Atlantic, the Canadian Arctic Archipelago and Baffin Bay, offers additional challenges but also other opportunities to theoretically estimate the transports. The geostrophic transports depend upon the number of openings (Stigebrandt, 1981) and two passages, Lancaster Sound and Nares Strait, transport Polar water from the Arctic Ocean to Baffin Bay. In addition, the West Greenland Current carries water, ultimately from Fram Strait and the Arctic Ocean, into Baffin Bay from the south. These inflows have to exit through one opening, Davis Strait, to the Labrador Sea.

The outflow from the Arctic Ocean is ultimately due to the higher sea level in the Arctic Ocean, which implies that the upper layer in Baffin Bay must be denser than that in the Arctic Ocean and its depth should be greater than the depth of inflows to Baffin Bay. There will be a pressure reversal in depth in Baffin Bay, below which the pressure gradient forces the Baffin Bay water to enter the inflow channels. The water from Baffin Bay does not enter the Arctic Ocean but it forms a lower boundary below the inflows that is taken as the depth of the transports to Baffin Bay. The depth and den-

sity of the Baffin Bay upper layer determine the geostrophic transport out of Baffin Bay, which should balance the northern inflows. The Arctic Ocean is connected to the deeper layers in Baffin Bay through Fram Strait and across the sills of the Greenland–Scotland Ridge and Davis Strait, removing any deep pressure gradient (Fig. 22).

Assuming that the temperature and salinity of the Baffin Bay upper layer are -1.6°C and 33.7, the upper layer depth in Baffin Bay can be found, and the geostrophic transports through the system can be determined as functions of the Polar water salinity or density (Rudels, 1986b). The transports decrease with increasing salinity of the Polar water, since the salinity difference between the layers is then reduced.

However, the characteristics of the Baffin Bay upper layer are not only due to the mixing between the entering waters but also influenced by the cooling and ice formation taking place in winter. The cooling and the ice formation only depend on the atmospheric forcing and are independent of the inflows, and a higher salinity of the entering Polar water would require an increased transport into Baffin Bay to reproduce the observed salinity of the Baffin Bay upper water.

Since one estimate gives a transport decreasing with increasing salinity and the other an increasing transport with increasing salinity, a salinity must exist for which these transports are equal. This salinity is then set as the salinity of the Polar water leaving the Arctic Ocean. The Polar outflow, the freshwater export, and the ice export through the Canadian Arctic Archipelago–Baffin Bay system can then be determined (Rudels, 1986b). However, the outflow estimate obtained by this approach, 0.7 Sv, was lower than the 2 Sv then commonly cited. One serious weakness of this approach is that it requires a residence time for the Polar water in Baffin Bay of at least one year for the transformation into the Baffin

Bay upper layer to take place. Especially the inflow through Lancaster Sound that enters close to Davis Strait may pass through Baffin Bay into the Labrador Sea in a time shorter than 1 year.

The expected change into a warmer climate has brought melting of the Greenland ice cap into focus. Intensified melting has been observed in the Jacobhavn isbrae in Baffin Bay, but also the low salinity East Greenland Coastal Current (Sutherland and Pickart, 2008), flowing on the Greenland shelf inshore of the East Greenland Current, indicates that freshwater is supplied from Greenland as runoff and as ice melt. This freshwater input, either directly to Baffin Bay and/or advected through Davis Strait into Baffin Bay, will lower the salinity of the upper layer in Baffin Bay.

The density contrast between the Polar water and the Baffin Bay upper layer then diminishes and the transport through the Archipelago becomes smaller. The expected increase in mass and freshwater fluxes from the Arctic Ocean to the North Atlantic in a warmer climate, might then not take place through the Canadian Arctic Archipelago. Instead, much of this increased outflow could be channeled through Fram Strait, which might affect the dense water formation and convection both in the Greenland Sea and in the Irminger and Labrador seas (Rudels, 2011).

It should be noted that in Rudels (2011) the ice formation and ice export are computed from the mass and salt balances in Baffin Bay using the postulated salinity of the Baffin Bay upper layer. It is then not independent of the inflow salinities. This in contrast to the approach in Rudels (1986b), where the ice formation was considered due to the atmospheric heat loss and unrelated to the salinities of the transports. It could then be used, together with the salinity of the upper layer, to determine the salinity of the Polar water entering Baffin Bay. The author appears, 25 years later, to have forgotten this aspect.

The Baffin Bay water column comprises not only transformed Polar water but also Atlantic water advected from the south and cold deep and bottom waters originating in the Arctic Ocean. These waters must eventually leave Baffin Bay, which requires a deep circulation in Baffin Bay that has not, so far, been considered. Rudels (2011) tried to theoretically estimate this deep circulation and how it relates to the upper layer exchanges by assuming that the inflows of upper water through the northern channels and in the West Greenland Current, here considered as part of the Polar inflow, are shallower than the depth of the pressure reversal.

This allows for inflows from the deeper layers to Baffin Bay. Furthermore, the pressure reversal between the Arctic Ocean and Baffin Bay does not occur in Baffin Bay but in the Labrador Sea, below the sill depth in Davis Strait. The pressure head due to the higher sea level in the Arctic Ocean is thus not removed in Baffin Bay and it forces not only a deep inflow in the north but also a deep outflow to the Labrador Sea. The pressure reversal in the West Greenland Current, by contrast, occurs above the sill depth and the deep Atlantic in-

flow recirculates in Baffin Bay and then exits through Davis Strait together with, but above, the outflow of the deep waters from the Arctic Ocean (Rudels, 2011).

The deep inflows decrease with increasing salinity of the Polar water, but the deep outflow through Davis Strait increases with the salinity of the Polar water. This implies that there is a salinity, 33.2 in this case, of the Polar water for which the deep exchanges in Baffin Bay are balanced. The deep outflow from the Arctic Ocean, taking place in the deepest channel, Nares Strait is 0.62 Sv, which is almost as large as the Nares Strait upper layer outflow of 0.77 Sv (Münchow and Melling, 2008). The added barotropic pressure gradient also increases the outflow of Polar water from the Arctic Ocean to the North Atlantic and the total outflow through the Canadian Arctic Archipelago–Baffin Bay system becomes about 2.2 Sv, somewhat larger than the about 1.5 Sv presently determined from observations (e.g., Curry et al., 2014). For details of this theoretical approach the reader is referred to Rudels (1986b, 2011).

13 The Subpolar gyre

All waters that circulate around the Arctic Mediterranean eventually flow south, entering the subpolar gyre in the North Atlantic. For most waters it is a return to where they started from as warm, saline Atlantic inflows crossing the Greenland–Scotland Ridge. Only the Pacific inflow and the freshwater input are added to the water entering the subpolar gyre from the north. We may remark here that in spite of the Pacific inflow bringing low salinity water to the Arctic Mediterranean, it constitutes a salt input to the Atlantic Ocean that eventually crosses the equator and is exported to the Southern Ocean.

The Atlantic water entering the Arctic Mediterranean becomes transformed in many ways by several different processes but, essentially, it is separated into denser and less dense waters by cooling and by receiving or rejecting freshwater, mainly through sea ice melting or sea ice formation. Waters that return to the North Atlantic either as dense overflows or buoyant outflows.

Østerhus et al. (2019) compiled the measured volume transports through all different control sections to find if the volume fluxes balance (Fig. 23 and Table 1). They found an inflow of 9.1 Sv and an outflow of 9.5 Sv, the exchange being almost in balance. However, an imbalance of 0.1 Sv would in 1 year lead to a sea level change in the Arctic Mediterranean of 0.2 m. This is more than that observed and indicates that the direct current measurements may still be improved.

However, what do not balance are the freshwater transports. Haine et al. (2015) estimated the freshwater storage in the Arctic Mediterranean in two periods, between 1980 and 2000 and between 2000 and 2010 (Table 2) and discussed the reason for the changes. What is clear, is that the adjustment time for the baroclinic pressure field is much longer than that

Table 1. Inflows, outflows and overflows in the different passages of the Arctic Mediterranean (based on Østerhus et al., 2019).

Branch and abbreviation	Period	months	Gaps Months	Transport (Sv)	Sd (Sv)
<i>Inflows</i>					
Denmark Strait, DS-inflow	Oct 1994–Dec 2015	250	5	0.9 ± 0.1	0.3
Iceland–Faroe, IF-inflow	Jan 1993–Dec 2015	276	0	3.8 ± 0.5	0.6
Faroe–Shetland, FS inflow	Jan 1993–Dec 2015	276	0	2.7 ± 0.5	1.1
European shelf, ES-inflow	Jan 1993–Dec 2015	276	0	0.6 ± 0.2	0.3
Bering Strait, BS-inflow	Aug 1997–Dec 2013	197	0	0.9 ± 0.1	0.4
<i>Overflows</i>					
Denmark Strait, DS-overflow	May 1996–Dec 2015	218	18	3.2 ± 0.5	0.4
Iceland–Faroe, IF-overflow	n/a			0.4 ± 0.3	
Faroe Bank, FB-overflow	Dec 1995–Dec 2015	206	35	2.0 ± 0.3	0.3
Wyville Thomson, WT-overflow	May 2006–May 2013	61	24	0.2 ± 0.1	0.1
<i>Surface outflows</i>					
Can. A. Arch., CA-outflow	Oct 2004–Sep 2010	72	0	1.7 ± 0.2	0.7
Denmark Strait, DS-outflow	Sep 2011–Jul 2012	11	0	2.0 ± 0.5	0.5
<i>Runoff and precipitation</i>					
Freshwater input, freshwater	n/a			0.2	

n/a: not applicable.

Table 2. Storage of liquid freshwater and sea ice in the Arctic Ocean and transports of liquid freshwater and sea ice through the main gateways of the Arctic Ocean (adapted from Haine et al., 2015).

Period	1980–2000	2000–2010
<i>Freshwater reservoirs (km³)</i>		
Liquid freshwater	93 000	101 000
Beaufort Gyre	18 500	23 500
Average sea ice cover	17 800	14 300
Total freshwater volume	110 800	115 300
<i>Freshwater fluxes (km³ a⁻¹ (m Sv)⁻¹)</i>		
Runoff	$3900 \pm 390/123.57 \pm 12.36$	$4200 \pm 420/133.08 \pm 13.31$
Bering Strait (liquid)	$2400 \pm 300/76.045 \pm 9.51$	$2500 \pm 100/79.21 \pm 3.17$
Bering Strait (sea ice)	$140 \pm 40/4.44 \pm 1.27$	$140 \pm 40/4.44 \pm 1.27$
$P - E$	$2000 \pm 200/63.37 \pm 6.34$	$2200 \pm 220/69.71 \pm 6.97$
Greenland flux	$330 \pm 20/10.46 \pm 0.63$	$370 \pm 25/11.72 \pm 0.79$
Davis Strait (liquid)	$3200 \pm 320/101.39 \pm 10.14$	$2900 \pm 190/91.89 \pm 6.02$
Davis Strait (sea ice)	$160 \pm ?/ 5.07 \pm ?$	$320 \pm 45/10.14 \pm 1.43$
Fram Strait (liquid)	$2700 \pm 530/85.55 \pm 16.79$	$2800 \pm 420/88.72 \pm 13.31$
Fram Strait (sea ice)	$2300 \pm 340/72.88 \pm 10.77$	$1900 \pm 280/60.20 \pm 8.87$
Barents Sea Opening	$90 \pm 90/2.85 \pm 2.85$	$90 \pm 90/2.85 \pm 2.85$
Fury and Hecla straits	$200 \pm ?/6.34 \pm +$	$200 \pm ?/6.34 \pm ?$
<i>Total fluxes (km³ a⁻¹ (m Sv)⁻¹)</i>		
Inflow sources	$8800 \pm 530/278.83 \pm 16.79$	$9400 \pm 490/297.85 \pm 15.53$
Outflow sinks	$8700 \pm 700/275.67 \pm 22.18$	$8250 \pm 550/261.41 \pm 17.43$
Residual	$100 \pm 900/3.17 \pm 28.52$	$1200 \pm 730/38.02 \pm 23.13$

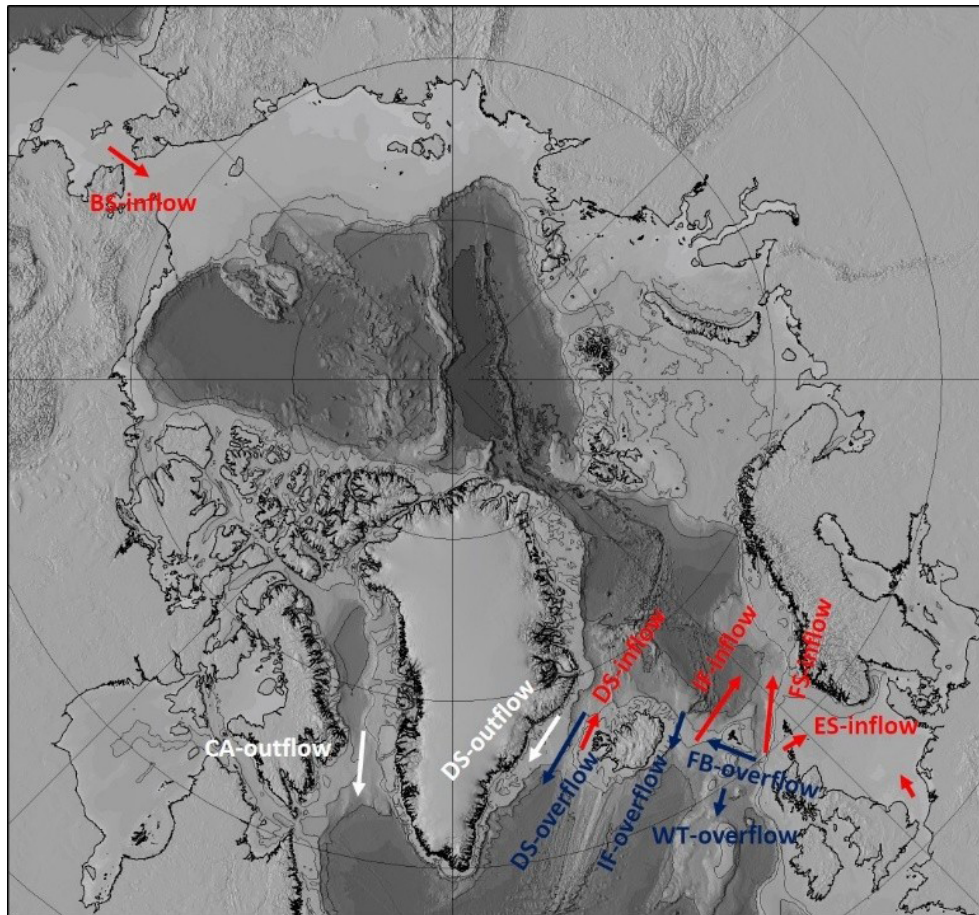


Figure 23. Map showing the volume balance of the Arctic Mediterranean Sea. BS: Bering Strait inflow, DS: Denmark Strait inflow, IF: Iceland-Faroe inflow, FS: Faroe–Shetland inflow, ES: European shelf inflow, CA: Canadian Arctic Archipelago outflow, DS: Denmark Strait outflow, DS: Denmark Strait overflow, IF: Iceland Faroe overflow, FB: Faroe Bank overflow, WT: Wyville Thomson overflow (adapted from Østerhus et al., 2019).

for the barotropic pressure field. Perhaps it will never really be in balance.

Of the Atlantic water about 6 Sv are transformed to dense overflow water while slightly more than 2 Sv become less dense water that together with the Pacific water make up the outflow waters (Østerhus et al., 2019). This is slightly more than the 1 Sv of less dense Polar water and the overflow of 1.8 Sv plus 3.2 Sv of overflow water estimated by Brown et al. (2025) and Bacon et al. (2022), assuming that the Atlantic water transformed in the Nordic Seas is dense enough to become overflow water.

The overflow water sinks into the deep and eventually joins the southward moving deep western boundary current, but the outflow water circulates in the upper part of the subpolar gyre and will influence the upstream conditions of the inflows to the Arctic Mediterranean.

The most conspicuous northern input to the subpolar gyre was the “Great Salinity Anomaly” observed in the late 1960s and early 1970s (Dickson et al., 1988). The outflow of low

salinity water and ice through Denmark Strait strongly reduced the salinity in the gyre and the anomaly could be traced around the subpolar gyre and then into the Nordic Seas at least as far as Fram Strait (Dickson et al., 1988).

Up to the mid-1990s the surface waters in the subpolar gyre were colder and less saline than their average temperature and the same conditions prevailed in the Nordic Seas and the Barents Sea. In the late 1980s to the mid-1990s the North Atlantic Oscillation (NAO) index, the normalized air pressure difference between the Azores and Iceland, was high and positive, and the cold continental air that dominated over the Labrador Sea led to a strong and deep convection and the created Labrador Sea water was exceptionally cold and fresh (Lazier et al., 2002). The cooling of the Labrador Sea accelerated the circulation of the subpolar gyre, which extended far to the east and partly prevented warmer water from the subtropical gyre to enter the Nordic Seas. The high NAO index was also manifested in the changes in the Barents Sea and in the Arctic Ocean, where the river runoff was diverted

farther to the east, from the Amundsen Basin to the Makarov Basin.

As the NAO index eventually became lower the subpolar gyre relaxed and retreated, leaving a freer passage for the waters of the North Atlantic Drift to reach the Greenland–Scotland Ridge and enter the Nordic Sea (Hátún et al., 2005; Häkkinen et al., 2008). After this a period of almost 15 year of warm, saline surface water was observed in the subpolar gyre and in the Nordic Seas and the Barents Sea.

The North Atlantic drift splits as it approaches the European continent and one part turns south and remains in the subtropical gyre. The simpleminded picture would be that it is the southernmost branch that turns south, but observations from floats launched in the upper 15 m indicated that most of the floats turned south and instead the deeper layers, around 300 m and deeper, enter the subpolar gyre (Brambilla and Talley, 2006; Burkholder and Lozier, 2011). On its path towards the Nordic Seas the water from the subtropical gyre is also cooled and freshened by mixing with the low salinity upper layers of the subpolar gyre flowing eastward from the Labrador Sea (Burkholder and Lozier, 2011).

In 2009 the RAPID Array at 26° N showed an increase of 2.5 Sv in the southward recirculation of the North Atlantic Drift (Smeed et al., 2014). This would reduce the warm water inflow to the subpolar gyre and less warm water would enter the Norwegian Sea. This change probably ended or at least contributed to the end of the 20 years period of warm surface water in the subpolar gyre (González-Pola et al., 2023).

In 2012 a strong freshening of the subpolar gyre was observed, the strongest in 120 years (Holliday et al., 2020). This freshening appears to be different from the one associated with the Great Salinity Anomaly in that it originated in a rerouting of the circulation of the low salinity Labrador Current. Instead of moving south around Newfoundland, it was detached from the continental slope and joined the subpolar gyre circulation eastward. This low salinity signal has continued northward, into the Barents Sea and to Fram Strait (González-Pola et al., 2023). If this freshening, taking place at least in the surface water, will imply a stronger stability of the Arctic Ocean water column and perhaps limit the vertical mixing in the Arctic Ocean and possibly affect the ice–ocean interactions, remains to be seen.

The atmospheric circulation patterns and the wind fields thus strongly influence the circulation into and out of the Arctic Mediterranean Sea, but are they also the main causes for the overturning circulation? The overturning requires that the northward flowing water becomes denser and sinks into the deep. Moreover, it also appears to require a sill that allows the convection to build up a reservoir of denser water and create a density difference between the deep waters on the two sides of the ridge. As the dense water crosses the ridge, the density difference allows it to accelerate and entrain ambient water, practically doubling its volume. This is different from the deep convection in the Labrador Sea that increases its depth and its density only gradually and marginally rela-

tive to the underlying waters. This might be the main reason for the small contribution from the Labrador Sea to the overturning circulation.

The formation of dense water in the Arctic Mediterranean also provides a constant, but perhaps weaker and often less obvious forcing than the wind driven circulation. The dense outflowing water has to be replaced, when the sea level starts to become lower, if not before. The outstanding question is: Can the processes in the Arctic Mediterranean produce enough dense, and dense enough, water in the future? If the convection does not reach to the bottom, it will create stagnant conditions in the deep basins, but as long as intermediate waters are formed, still denser than the waters in the subpolar gyre, and the pressure head at the sill is present, the overturning circulation should be maintained. A large input of freshwater to the Greenland Sea, either from the Arctic Ocean or from ice melting on Greenland, could strongly affect the convection in the Greenland Sea (Stigebrandt, 1985). Only the cooling of the Atlantic water in the Norwegian Sea and the unknown rate of deep water formation in the Arctic Ocean would then remain as sources of dense overflow water, but that might be sufficient for maintaining the overturning circulation.

14 Overview and Global connections and interaction

At the global scale the Arctic Ocean is a fundamental node in the global hydrological cycle and the ocean's thermohaline circulation (Fig. 24). Here we have assessed the system's circulation, water mass transformations, and key functions and processes. Our use of the term *Arctic Mediterranean* acknowledges that the Arctic Ocean and Nordic Seas are largely surrounded by north-flowing drainage basins supplied by moisture originally derived from subarctic seas and delivered by the Westerly wind field (cf. Carmack et al., 2008, 2016). As such they are termed *beta oceans*, where stratification is controlled by salinity which, in turn, allows ice formation and attendant seasonal fluxes of salt and freshwater (Carmack, 2007). The Arctic Mediterranean is, itself, dynamically and thermodynamically, a *double estuary*, in that its inflowing subarctic waters are transformed into outflow waters that are both lighter and denser than their parent water masses (Carmack and Wassmann, 2006; Rudels and Carmack, 2022; Brown et al., 2025), all flowing along the east and west coasts of Greenland. There they mingle and mix with new incoming subarctic waters to assemble the lower limb of the Atlantic Meridional Overturning Circulation.

The Arctic climate system, is now primed to cross multiple cascading and irreversible tipping points, as was forecast by Aagaard and Carmack (1989) who speculated, “that the present-day Greenland and Iceland seas, and probably also the Labrador Sea, are rather delicately poised with respect to their ability to sustain convection.”

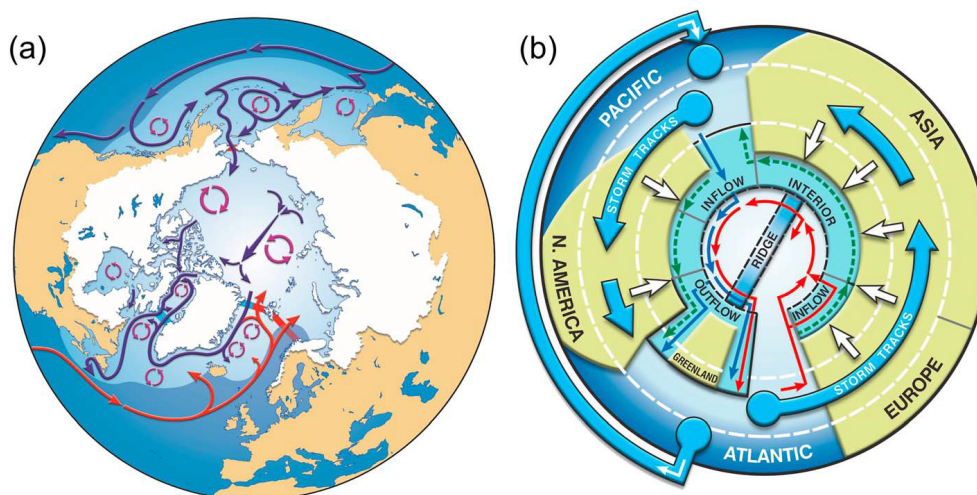


Figure 24. Schematic maps of (a) the major ocean currents (long arrows), the four Arctic Ocean gateways in Fram Strait, the Barents Sea Opening, Davis Strait, and Bering Strait (thick arrows with red denoting warm and saline inflow and blue denoting cooler and fresher surface flows), the gyral circulation patterns (circular arrows), the salt-stratified ocean domains are shown in light blue, and the terrestrial regions contributing freshwater to the Arctic Mediterranean shown in white, and (b) a highly stylized summary of components of the high-latitude freshwater system. Here moisture is transported from the subtropical and tropical Atlantic Ocean to the Pacific Ocean via the Trade Winds over Central America (thick blue arrow). The subarctic front separates the thermally stratified subarctic oceans (darker blue) from the salt-stratified Northern Ocean (lighter blue) in both the Atlantic and Pacific Oceans. Moisture is transported from the Pacific and Atlantic Oceans to the Arctic catchment basins by the midlatitude (Westerlies) storm tracks (thick blue arrows), which subsequently drains into the Arctic Ocean (thick white arrows) where it spreads initially within coastal domains (dashed green arrows). Warm, salty Atlantic origin waters (thin red arrows) enter the Arctic Ocean through Fram Strait (the Fram Strait Branch) and through the Barents Sea Opening (the Barents Sea Branch) and circulate within the Arctic Basins as subsurface, cyclonic, topographically steered boundary currents along the continental margin and ridge system. Internally modified Atlantic waters exit the Arctic Ocean southward through Fram Strait along eastern Greenland. Cooler and fresher Pacific origin waters (thin blue arrows) enter the Arctic Ocean through Bering Strait and exit through the Canadian Arctic Archipelago and Fram Strait along eastern Greenland. Within the Arctic Ocean a topological distinction is made between inflow, interior, and outflow shelves (cf. Carmack and Wassmann, 2006; Bluhm et al., 2015).

Data availability. The figures in this work have all but two (Figs. 8 and 11) been published before (see the figure captions). The figures are based on calibrated hydrographic data obtained directly from the cruises, before they were sent to some data center, where the data now are available.

Author contributions. BR received and accepted the invitation to submit an article to the OS 20 years Jubilee volume. He suggested the title and planned the structure and scope of the article and asked EC to join in the writing, knowing that a combination of expertise would make for a better article.

Competing interests. The contact author has declared that neither of the authors has any competing interests.

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