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The circulation and water mass (trans)formations in the Arctic Mediterranean Sea and their impact on the ocean deep circulation: a review.

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

Summary

This review traces the pathways and transformations of water masses within the Arctic Mediterranean Sea, and elucidates how waters leaving this domain differ from those entering. Processes constraining sea ice formation and melting due to both atmospheric interaction and heat and salt exchanges with the underlying ocean are central to understanding the present and future Arctic Ocean. The focus is on the Atlantic water, how its northward circulation affects the Arctic conditions, and how returning, transformed water impacts the global circulation.

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Section 1 Introduction

The mean surface temperature of the earth is rising and how this may 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 primarily are covered by ocean and one question is: Does the presence of an ocean accelerates or mitigates the observed changes? To address this and related questions we review in section 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 bathymetric features of the Arctic Mediterranean Sea 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 has elapsed since the exploration of the northern latitudes started. The section also reviews the circulation and the water mass (trans)formations taking place in the Arctic Mediterranean, and how our understanding has 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 establishment of the Committee for the Exploration of the Sea (ICES) in 1899 and 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

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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 the wind driven model proposed by Nøst and Isachsen (2003) is also discussed.

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 Siberia and as the direct inflow of low salinity Pacific water through the shallow Bering Strait (section 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 double-diffusion and diffusive interfaces. In the Transpolar Drift the Siberian branch transports runoff and sea ice from the Laptev Sea and the anticyclonic Beaufort Gyre brings low salinity water towards Greenland. The convergence of the streams leads to exchanges between the two gyres. The differences between the deep waters in the Amerasian and Eurasian basins and their causes are also described and discussed.

In section 9 the exchanges through Fram Strait are described from two hydrographic sections, one north of the strait and one along the sill. The different waters are identified and the transports of volume, freshwater and heat through the strait are discussed based on the geostrophic estimates obtained by Rudels et al. (2008).

The Nordic Seas, especially the Greenland Sea, are regions characterized by weak stability and are areas for deep convection and deep water renewal and the convection in the Greenland Sea is discussed in section 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 process has changed. This is perhaps the area, where the effects of a warming climate have been most evident.

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

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section 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 occur through Fram Strait in the East Greenland Current and through the channels in the Canadian Arctic Archipelago to Baffin Bay are discussed in section 12. As the East Greenland Current passes Cape Farewell it turns north as the West Greenland Current and also enters Baffin Bay. The outflows 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, where the transports are estimated from geostrophic exchanges through the channels and the 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 also discussed. Finally, the exchanges between the Arctic Mediterranean and the subpolar gyre and the Labrador and Irminger seas are discussed in section 13, where also the impact of water mass transformations taking place in a semi-isolated sea could have on the overturning circulation.

This review is written with focus on observation and their interpretation. Processes are described and their possible importance discussed, occasionally by the use of conceptual models. For further aspects of the oceanography of the Arctic Mediterranean we refer the readers to the review articles by Coachman and Aagaard (1974), Timmermans and Marshall (2020), Rudels and Carmack (2022), and the books by Lewis et al., (2000), Dickson et al. (2008), Mills (2009) and Rudels (2021).

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Section 2 The 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. 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^{o} to the ecliptic. This leads to large seasonal variations in incoming solar radiation especially north and south of the polar circles (>66.5^{o}). In summer the incoming radiation at the top of the atmosphere is larger at the poles than at the equator, while in winter the sun is below the horizon and the long, dark polar night reigns. The contrasts between the two polar regions are, however, large. 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^{o} to 25^{o} N(S). Some of the sinking air returns towards the equator, creating 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 and returns equatorward forcing the warmer air of the Ferrel cell to rise.

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At low latitudes the poleward oceanic heat transport is larger than the corresponding atmospheric transport, but the transports become equal around 30°N(S). At higher latitudes the ocean transport diminishes and beyond 65° only the heat transport in the North Atlantic is different from zero. The Atlantic is different from the other oceans in that it carries heat across the equator from the southern to the northern hemisphere. The warm water ultimately comes from the Indian Ocean and it 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 condensates, 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 large freshwater input, 10% of the entire global continental runoff, to the Arctic Ocean.

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 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 temperature close to melting temperature as long as ice is present. Ice formed and melted does not contribute any net heat to the Arctic Ocean, but the

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melting in summer stabilizes the water column and isolates the deeper layers from the surface forcing both in summer and 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 stagnant North Pacific and the ventilated North Atlantic.

The Arctic Ocean is the smallest of the Earth's oceans ($9.5\ 10^{12}\ m^{12}$), 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 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 (Figure 1).

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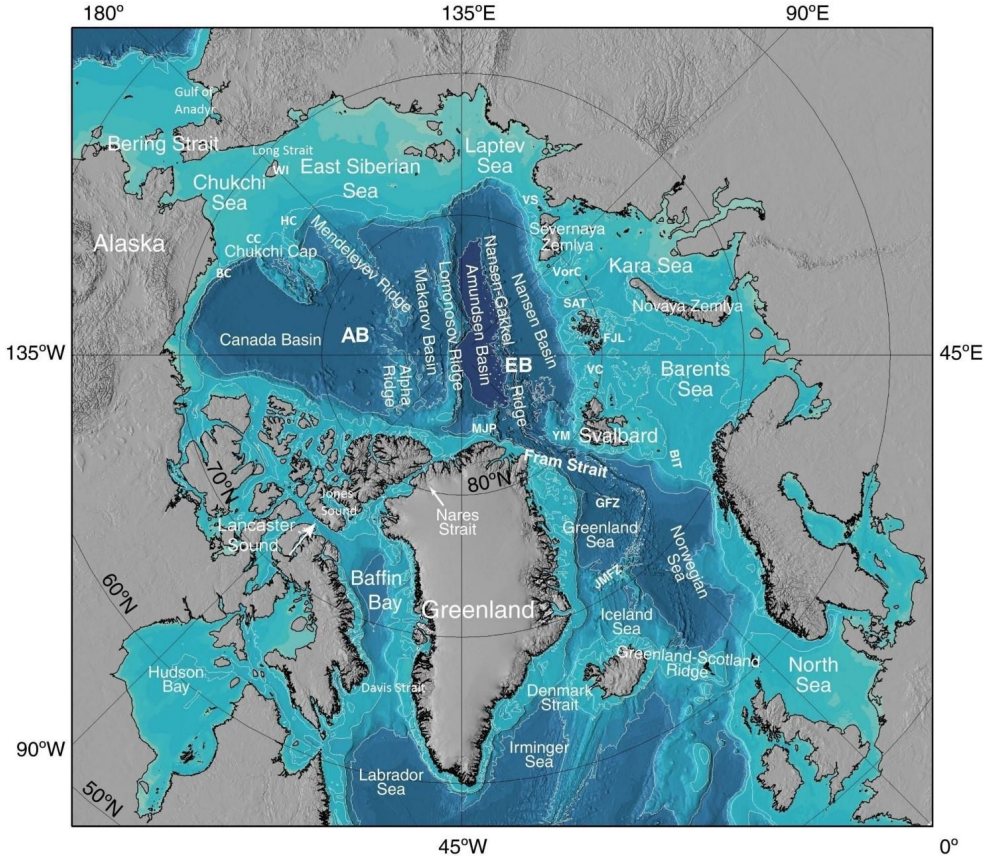

Figure 1: Map of the Arctic Mediterranean Sea showing geographical and bathymetric features. The bathymetry is from the IBCAO updated data base and the projection is Lambert Equal Area. The 200, 500, 2000 and 4000 m isobaths are shown. AB; Amerasian Basin, BIT; Bear Island Trough (Barents Sea opening),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

The deeper part of the Arctic Ocean consists of major two basins, the Amerasian and the Eurasian Basin, separated by the narrow Lomonosov Ridge with an average sill depth around 1600 m. The Amerasian Basin is divided by the Alpha and Mendeleyev 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

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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 Norwegian 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 across the Greenland-Scotland Ridge connect the Nordic Seas to the North Atlantic, the 640 m deep Denmark Strait between Greenland and Iceland, the 400 m 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 m 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, 45 m.

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Section 3 Exploration

Section 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 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 Frenchman John Cartier explored 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 it was no passage to the Pacific. The English search eventually concentrated on higher latitudes. Martin Frobisher 1576 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 but on his return voyage he noticed a strait with strong current

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that might lead to a passage west. In 1610 Henry Hudson was given the task to examine this passage, but after passing through the strait, later named after him, and entering the large open water area autumn was approaching and 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 facing a certain death.

When the ship returned to London the crew presented Hudson's sailing notes and was pardoned. Further expeditions showed that Hudson had sailed into a Bay, Hudson Bay, not into a passage to the Pacific. In 1616 Robert Bylot, with William Baffin as pilot 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 they were not leading to the Pacific. Baffin's sailing notes were, unfortunately, 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 Barents 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 Barents died already in 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

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283 independent from Spain and Spain had given up its monopoly of the southern routes to Cathay. No
284 pressing need for the Dutch to find the Northeast Passage then remained (Figure 2).
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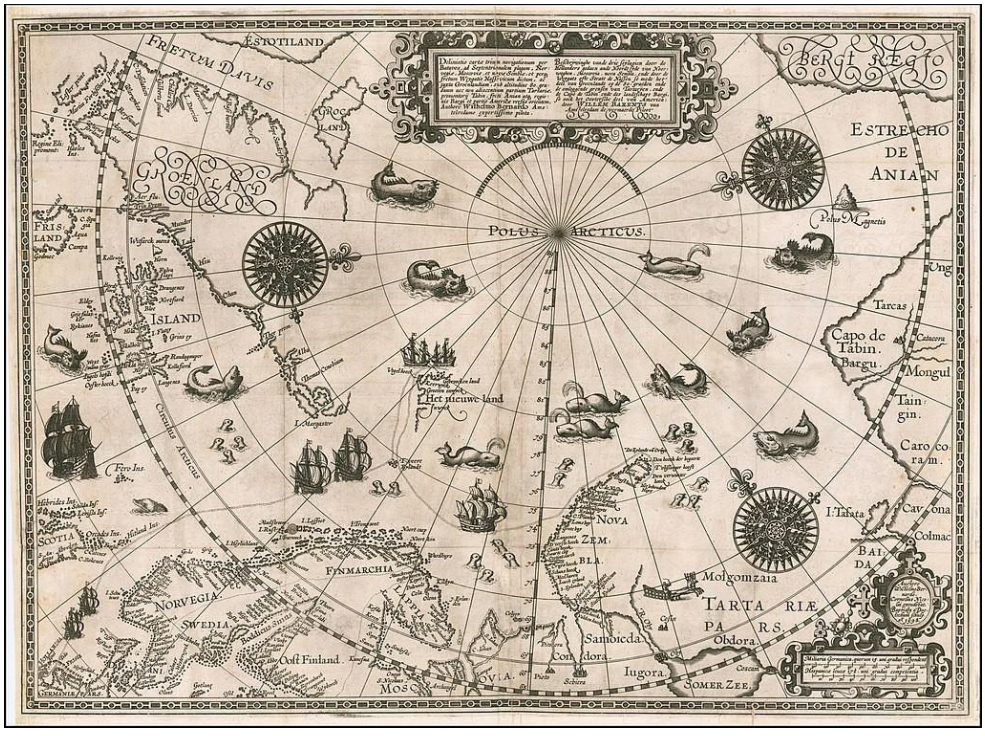

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288 Figure 2: A map of the Arctic Ocean by Willem Barents drawn around 1598 showing the then known
289 features of surrounding lands and islands. The ships north of Svalbard probably indicate open water,
290 the Whalers' Bay (from Häkli, 1992).
291

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

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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 to extend 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 on his first expedition observed, and on his second expedition reached Alaska, but he 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 had 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 already 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 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 mapmaker 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.

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3.2 The deep circulation

In the age of sail the separation of the direct driving by the wind from the effects of the ocean currents was not easy to make, and 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, two 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* and supported the 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*". 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 Transaction 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 (1797) also pointed out that ocean water does not attain its maximum density at 4°C, 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 self-closing buckets on Otto von Kotzebue's second circumnavigation in 1823-1826, observing 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) (Figure 3).

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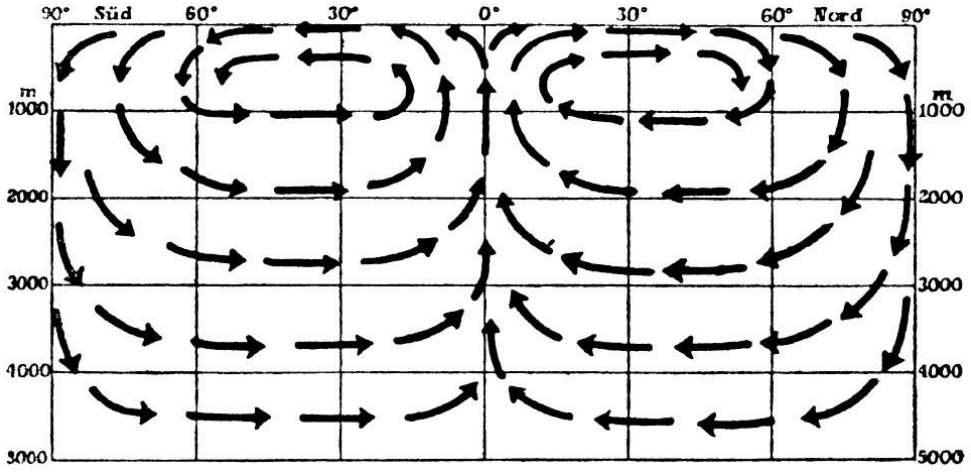

Figure 3: The symmetrical thermally driven meridional overturning circulation suggested by Emil von Lenz 1845 (from Merz and Wüst, 1922).

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 exist at the ocean bottom regardless how deep they sampled. However, they also found deep water with temperatures between 0°C 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°C 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

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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 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

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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 -1.2C^{o} 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

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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 (1877) tried to convince the Dutch authorities to revive the Dutch Arctic explorations, which had been so intense during the 17th century, by sending expeditions following different warm currents into the Arctic Ocean.

However, there were also contrary views. Maury (1855) 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 be a circulation akin to that suggested by von Weitz (1755) (see below).

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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 1972 with the ship *Admiral Tegetthoff* towards the Arctic Ocean east of Svalbard, and Charles 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 Frans 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 two 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

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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 of 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 with 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 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 enough 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 at 150 m to 200 m

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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 drift of *Fram* did not, however, remove the 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. Planning for a submarine expedition to the North Pole with a discarded submarine *Nautilus* was made in 1931 but the expedition had to be abandoned just as it had started because failures with the diving rudders, preventing the submarine to dive (Sverdrup and Soule, 1933). The Soviet ice breaker *G. Sedov* was trapped in the ice in the late 1930s 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 known. Observations of the density distribution in the water column also showed that the wind field creates an anticyclonic geostrophic

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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 (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. 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, really 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*. However, the report on these expeditions, *The Norwegian Sea,* published by Bjørn Helland-Hansen and 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.

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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. 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 at 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 be present also farther south (Nansen, 1913). The other 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 *Ingolf* 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 task that actually was achieved 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, 1898b, 1901). How to use the circulation theorem to compute

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ocean currents was described by Sandström and Helland-Hansen (1903) and Helland Hansen (1905). 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 Greenland 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 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.

Sandström (1908) made laboratory experiment with saline water in a narrow tank to describe the thermally driven circulation in the North Atlantic (Figure 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

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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.

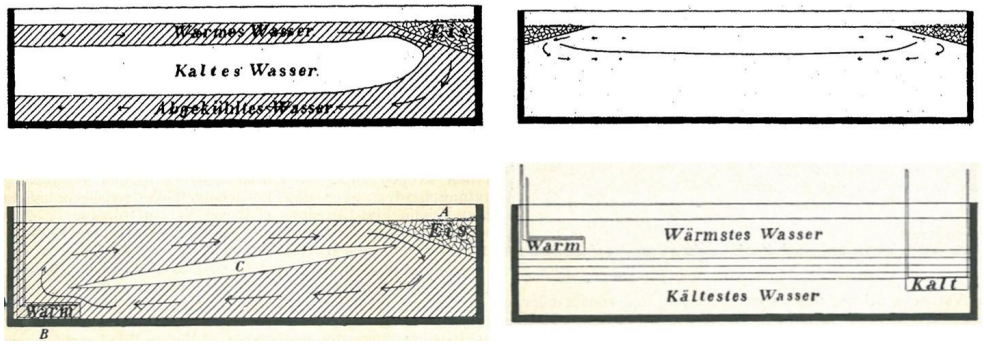

Figure 4 The Sandström experiment to simulate the circulation between lower and higher latitudes driven by cooling at high latitudes. The upper panel shows the effect by just cooling, at the northern hemisphere (left) and at both hemispheres (right). The circulation is sluggish. The lower panel shows 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 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. 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 Bosporus studied by Luigi Marsigli (1681), which showed less saline water from the Black Sea flowing

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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 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.

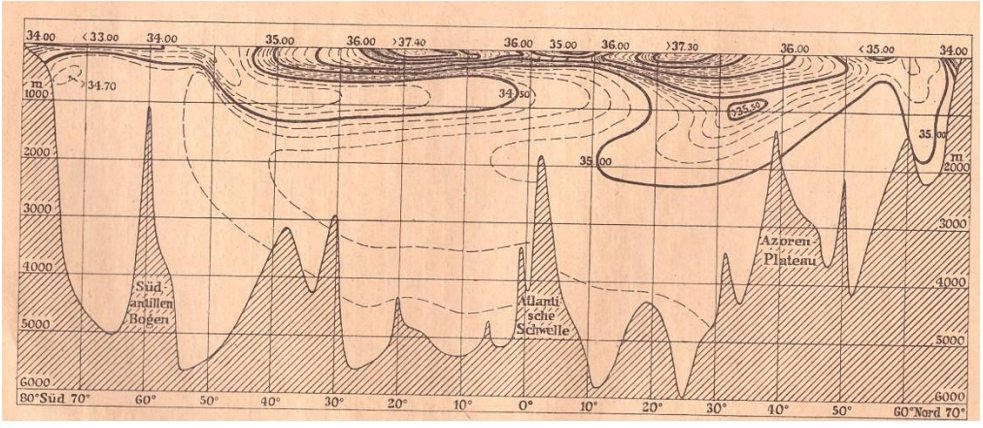

Figure 5. A longitudinal salinity section in the Atlantic showing the present of the upper low salinity input from the Southern Ocean, the saline North Atlantic deep water and the slightly less saline bottom water from the Antarctic (from Merz and Wüst, 1922).

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 temperature and salinity influenced the circulation. Merz and Wüst 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 m to 4000 m depth a more saline and colder layer flows south from the

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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 (Figure 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.

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Section 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, mainly as strongly seasonal runoff from the large catchment areas but also as direct local precipitation.

The Pacific water that enters the Arctic Ocean through Bering Strait is driven by the higher sea level in the North Pacific than in the Arctic Ocean (Stigebrandt, 1984). The sea level difference is caused by the less dense water column in the North Pacific 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 transports in winter have been explained by the presence of stronger, more northerly winds (Coachman et al., 1975). The inflow in summer is warm, creating the Bering Strait summer water, while the inflow in winter is at the freezing temperature forming the Bering Sea winter water, showing up as a temperature maximum and a deeper temperature minimum respectively in the Canada Basin water column. The Pacific inflow is limited by the shallow depth (85m) of the strait, and it does not carry any oceanic heat from lower latitudes. The warmer water transported in summer is formed by local seasonal heating and is rather a part of the radiation balance of the Arctic Mediterranean. Because of its low salinity the Pacific inflow also transports a substantial amount of freshwater to the Arctic Ocean, 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 mean sill depth of about 500m. The thickness of the Atlantic water layer entering the Norwegian Sea is 300-400m 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

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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 connecting 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. This recirculation of Atlantic water was 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 Scotland it splits. The upper layer moves mainly south and remains in the subtropical gyre, while the deeper layers, below 300m, raise 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 affect the properties of

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the northward flowing Atlantic water is not adequately known but likely depends on 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 would have a large impact on how the Arctic Mediterranean responds to changes in climate and we return to this question later.

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Section 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 upper 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 entire water column has to be cooled to freezing temperature or down to a density gradient caused by an advective input of a denser water mass. As ice starts to form above a water column at the freezing point no 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 the perhaps 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

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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).

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 ice and brine. 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 to be 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

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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.

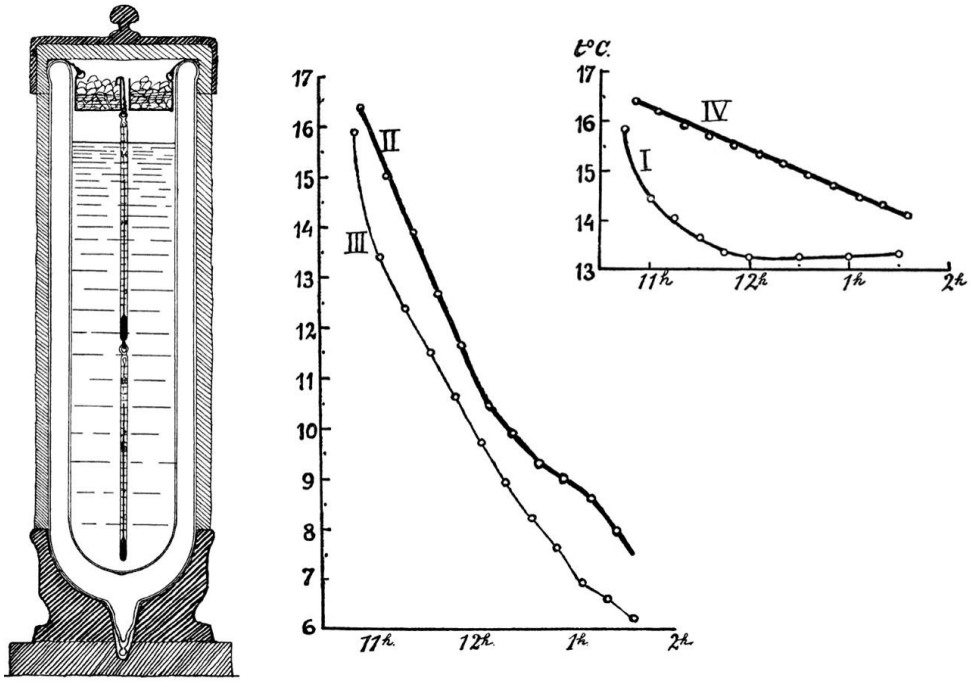

Figure 6. Nansen's experiments with cooling of sea water. Experiment I is when the ice is placed directly on the water, which leads to the formation of a low salinity melt water layer and no or little convection in and cooling of the water column. Experiments II and III show the results when the ice 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 (from Nansen, 1912).

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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 water across the interface into the upper layer. Furthermore, in summer the main sea ice melting occurs at the surface and the melt water prevents heat from below to reach the ice. In winter, however, the air temperature is low and the summer melt water layer is gone. 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 then becomes:

$$w_e = \frac{2n_o u_*^3}{g[\beta(S_A - S_1) - \alpha(T_A - T_1)]} - \frac{\varepsilon B}{g\left[[\beta(S_A - S_1) - \alpha(T_A - T_1)]\right]} \quad (1)$$

Here n_o is a constant usually put to 1.25, u_* is the friction velocity, T_A, T_1, S_A and S_1 are the temperatures and salinities of the upper and lower layer respectively, α, 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

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fraction of heat going to ice melt is the one that leads to minimum ice melt (Rudels et al., 1999a) This fraction ϕ_o is found to be:

$$e\phi_o = \frac{2\alpha L}{c(\beta S_A - \alpha \Delta T)} \approx \frac{2\alpha L}{c\beta S_A} \quad (2)$$

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 Figure 7 shows the total heat loss Q', the heat loss to ice melt $\phi_o Q'$, the heat loss to the atmosphere $(1-\phi_o)Q'$ as functions of ϕ for different ΔT and the corresponding values of α.

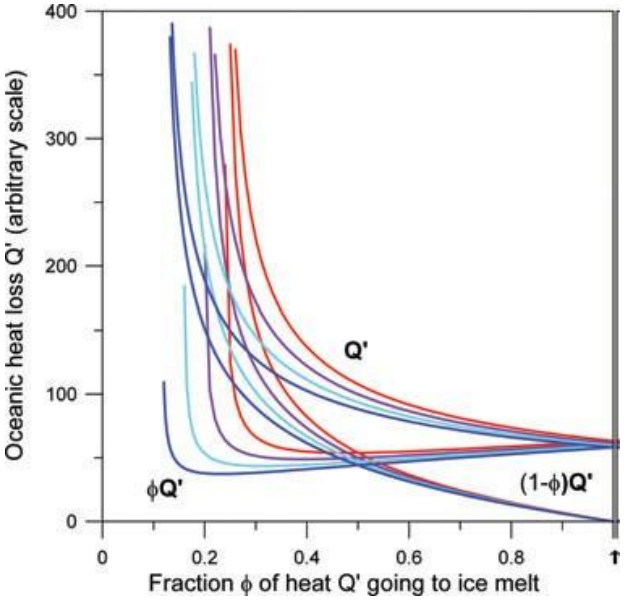

Figure 7. Curves showing the total heat loss Q', the heat going to ice melt $\phi Q'$ and the heat going to the atmosphere $(1-\phi)Q'$ as function of ϕ, the fraction going to ice melt. Red curves have $\Delta\alpha T = 8°C$ and $\alpha = 0.781 \times 10^{-4}$, purple curves $\Delta\alpha T = 6°C$ and $\alpha = 0.654 \times 10^{-4}$, cyan curves $\Delta\alpha T = 4°C$ and $\alpha = 0526 \times 10^{-4}$, blue curves $\Delta\alpha T = 2°C$ and $\alpha = 0.390 \times 10^{-4}$. $\beta = 8 \times 10^{-4}$, c 4000Jkg^{-1}K^{-1}, L = 0.335 x 10^{6}Jkg^{-1}. The fraction ϕ_o giving minimum ice melt decreases with decreasing temperature and decreasing α.

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825

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

829

830 $$S_1 \approx \frac{S_A}{\left(1+\frac{2\alpha}{\beta S_A}(T_A - T_f)\right)} \quad (3)$$

831

832 T_A is the temperature of the lower layer and T_f the freezing temperature of the upper layer. The
833 stable salinity step $\beta(S_A - S_1)$ is about twice the unstable temperature step $\alpha(T_A - T_1)$ giving a stability
834 ratio of 2 at the interface. The problem is the choice of α, which depends strongly on temperature
835 and as the upper layer deepens and cools α decreases to a minimum at the freezing point. A smaller
836 α leads to a deeper upper layer with higher salinity.
837 When the freezing temperature is reached, the upper layer has become so deep that
838 the amount of entrained warm water is not enough to supply heat both to ice melting and to the
839 atmosphere and ice again starts to form. The stability at the interface then decreases and
840 entrainment starts to increase. Whether this leads to a stable situation with a deepening of the
841 upper layer, more melting and a thinning of the ice cover, or to an increase in salinity and density of
842 the upper layer eventually leading to deep convection will be discussed in section (10) below. Here
843 we might just add that the fraction ϕ_o of minimum heat going to ice melt is twice the fraction leading
844 to marginal stability. A fraction smaller than $\phi_o/2$ will not stabilize the water column and it overturns.
845 This is also evident from the stability ratio $R_\rho = \beta\Delta S/\alpha\Delta T = 2$ reached when the upper layer has
846 attained freezing temperature.
847

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Section 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. 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, potentially contributing to the overturning circulation, instead will leave the Arctic Mediterranean as less dense upper water. 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 shortwave 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 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 500m deep Victoria Channel west of Franz Josef Land. The northern

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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 passage 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 also penetrates northward onto the Grand Bank.

In the 2010s the ice cover in the northern Barents Sea became 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 et al., 2012; Lind et al., 2018). The change would be part of a larger shift that is presently occurring in the Arctic Ocean, the “Atlantification” of the Arctic Ocean, which implies a reduction of the upper layer thickness and stability and a stronger upward transport of heat from the Atlantic layer to the ice (Polyakov et al., 2017). 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). 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 i980s.

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

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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 1980 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. 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).

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Section 7 The circulation in the Arctic Mediterranean Sea

Section 7.1 Atlantic water in the Eurasian Basin

The two Atlantic inflow branches meet again 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 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 covering the Fram Strait branch.

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, 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 at 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

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upper part of the water column. The Atlantic water becomes more isolated from the ice cover, the sea surface, 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.

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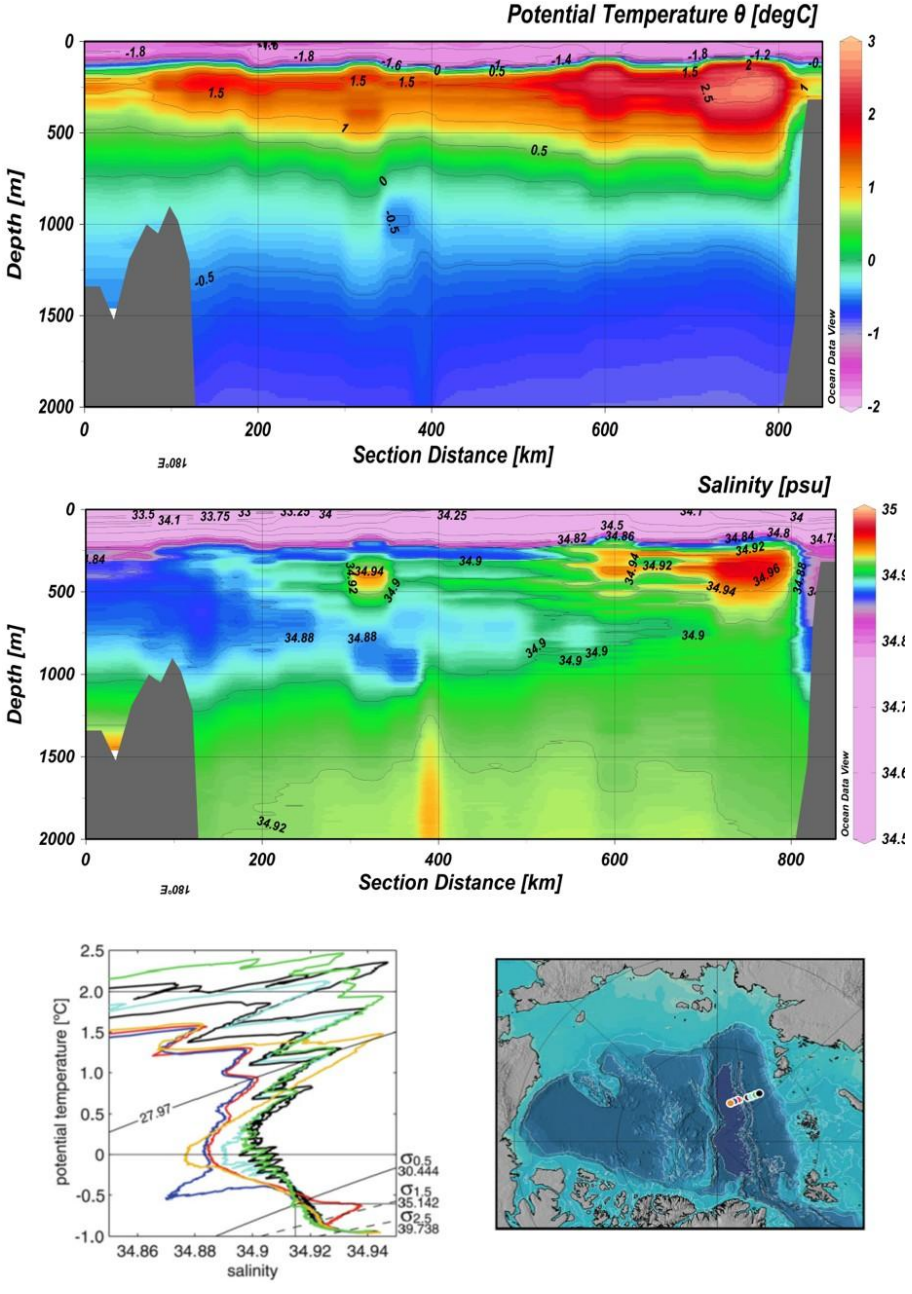

Figure 8. Potential temperature and salinity sections taken 1996 by *FS Polarstern* from the eastern Kara Sea shelf across the Nansen and Amundsen basins and the Lomonosov Ridge into the Makarov Basin. The θS curves are from the same section, showing interleaving as well as eddies in the different waters.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

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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 Giorgi, 1981). However, the intrusions penetrate into the opposite water mass only as long as an unstable distribution of the driving component remains. Once it is removed, the interleaving stops. 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 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 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. Instead, 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 a heat loss to the overlying water

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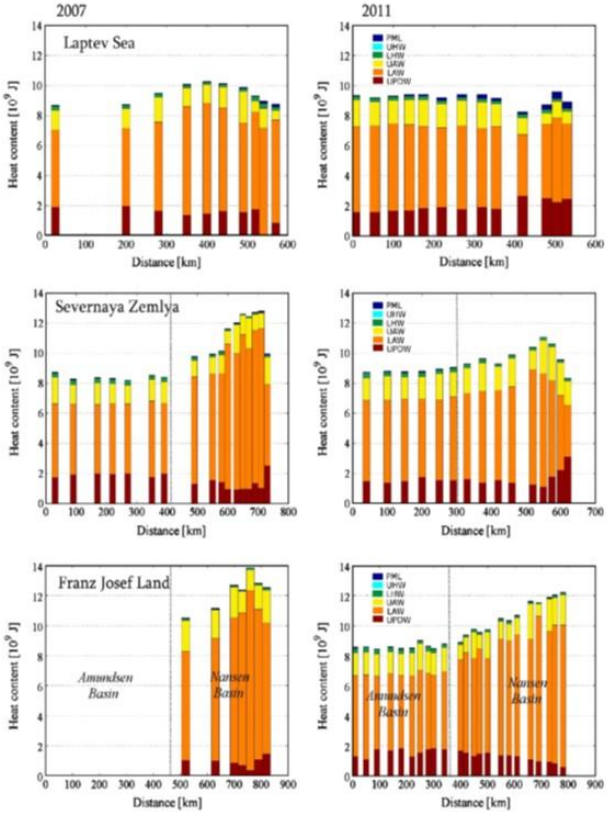
and to the ice and atmosphere. The heat has already been lost elsewhere, mainly in the Barents Sea. Not only is the Atlantic core cooled, the salinity also decreases significantly (Figure 10). To mix freshwater or low salinity water mechanically and vertically into the deep is energy consuming and the 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 in the surface layer (Fig. 9). To understand how present, and especially how possible future changes in inflow strength and characteristics of the Atlantic water might influence the future condition in the eastern Nansen Basin, it is necessary to consider both inflow branches, how they are transformed individually and how they interact.



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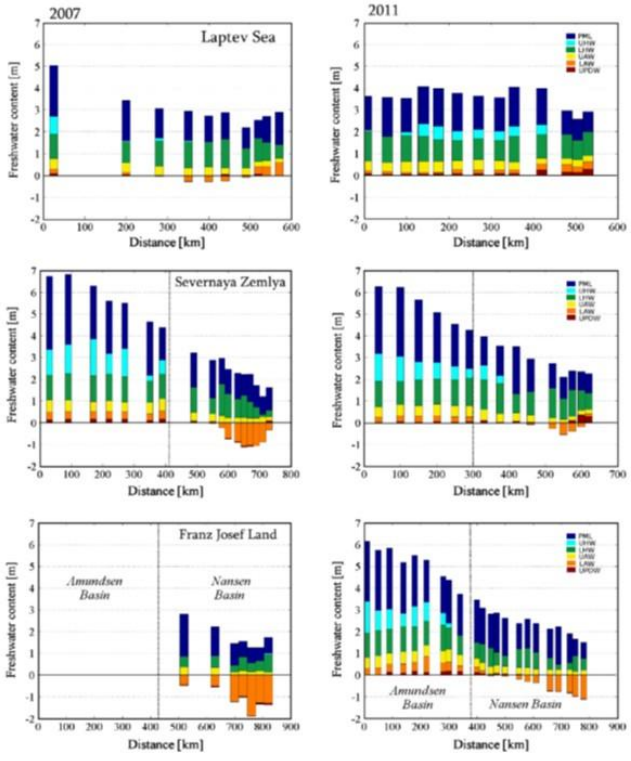

Figure 9. Heat (upper panels) and freshwater (lower panel) content in the upper 1000 m of the water column (the height of the columns, except where the freshwater content is negative in the Atlantic water) and the different water masses observed on sections taken in the Eurasian Basin in 2007 and 2011. 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 θ<0°C and 34<S, Atlantic water I (AWI) 0°C<θ, and above the temperature maximum. Atlantic water II (AWII) 0°C<θ, but below the temperature maximum, upper Polar Deep Water θ<0°C, below AWII and above 1000 db. Especially the increase in freshwater content (reference salinity 34.9) 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 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

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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 interleaving into a thermohaline step structure with thick homogenous 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., 1993). 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 (Morison 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

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a salinity minimum between 800m and 1200m in the Eurasian Basin (Rudels et al., 2013, see also figure 8). In the 2000s, by contrast, the water that enters and passes over the Barents Sea to the Arctic Ocean has been warmer, more saline and denser and has instead been identified as a salinity maximum around 2000 m (Rudels, 2021).

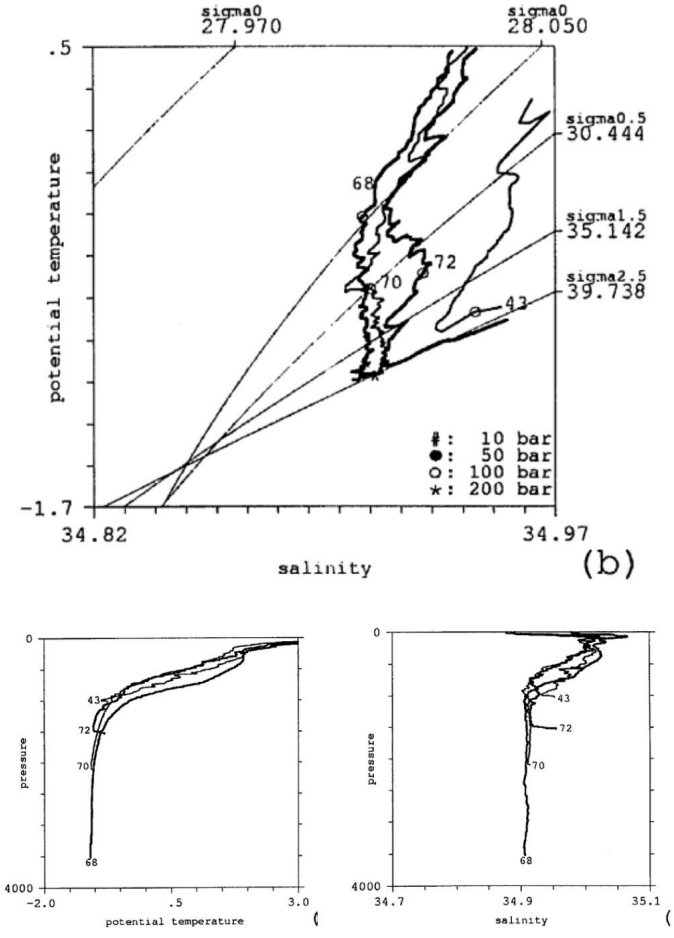

Figure 10. Potential temperature and salinity profiles and θS curves observed from FS Valdivia west of Svalbard in 1986. The increase in salinity and temperature in the deepest layer indicates brine enriched water from Storfjorden that has entrained intermediate warm Atlantic water during its descent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

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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., 2001). 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 the descending plumes. The 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 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 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 t. Anna Trough, and perhaps also the dense overflow plumes crossing the Greenland-Scotland Ridge. The entraining plumes, by contrast, would be thin and carry but little water from the shelf but instead incorporate intermediate water from 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 observations shown in Fig. 10 suggest entraining plumes.

Section 7.2 A possible wind driven circulation

The two Atlantic inflows partly cross the Lomonosov Ridge and enter the Amerasian Basin, partly

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remain 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 and several conceptual models have been proposed and here the approach taken by Nøst and Isachsen (2003) will be examined.

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 suggested either quadratic or linear bottom friction laws to represent the bottom torque. The applied meteorological forcing corresponds to a vertical velocity of about $10^{-6}ms^{-1}$, 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 $10^{-1}ms^{-1}$, which is close to the observed bottom velocities (Newton and Sotirin, 1997; Nøst and Isachsen, 2003). The picture is then 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 Nost (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

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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 push the boundary current to deeper levels and replace 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 the 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 an interesting possibility.

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Section 8 The circulation in the Amerasian Basin

East of the Laptev Sea and the New Siberian Islands the nature of the circulation changes. It is no longer characterized by the two Atlantic inflows but by freshwater input from rivers and from the low salinity Pacific inflow. 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 rim 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 (Morison 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*, 1918-1025, 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 located west of the New Siberian Islands. while the East Siberian Sea, in contrast to the Laptev Sea, more acts as an ice sink for the inner Arctic Ocean. This is partly confirmed by Zacharov (1976), who estimated the ice

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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. The decline in ice cover 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 m to 1.8 m (Rothrock et al., 1999).

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 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 4500km^3 and the liquid freshwater that

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has accumulated in the Beaufort Gyre has increased the freshwater storage there by 5000km^3, between 2000s and the 1980s and 1990s. This more than compensate for the about 3500km^3 loss of sea ice (Haine et al., 2015).

The circulation in the Beaufort Gyre is complex. The anticyclonic winds collect the low salinity upper water in the center of the gyre and 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 disappear, 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. One, albeit speculative, possibility could be a Sverdrup balance. The anticyclonic wind field generates a southward transport, which is balanced by an inertial return flow on the western side of the gyre, like the Transpolar drift. The flow would intensify as long

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as f increases, but when the highest latitudes is passed f starts to decrease and the boundary current begins to diverge. This is but speculation.

In the Amerasian Basin the input of low salinity water from the shelves, and the Pacific inflows that create a temperature maximum around 75m in summer and form a temperature minimum between 200m and 300m in winter, further isolate the Atlantic water from the energy input at the sea surface. The temperature minimum also involves an increase of salinity, suggesting that in winter ice has formed in the Bering and Chukchi seas, and 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 regeneration of nutrients 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 and steps appear to be coherent over several kilometers, giving a horizontal to vertical ratio of about 10000 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 in time rather than in space.

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

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Atlantic water enters the Amerasian Basin and the larger number of layers and interfaces compared to the Eurasian Basin would be due to a decreased buoyancy 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.

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 the 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 implied that the deep Arctic Ocean was not stagnate 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

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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 Ocen 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 m to 2000 m the deep waters in the different basins communicate less freely and they have acquired their different characteristics. 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 m to 800m thick bottom layer is encountered. The temperature increase and the homogenous bottom water have been explained by geothermal heating, which could homogenize

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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). The presence 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 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 continues to decrease until 500 m from the bottom. Below this level a homogenous bottom layer is found. This feature was explained by Jones et al. (1995) by a spillover across the Lomonosov Ridge of colder, less saline deep water from Amundsen Basin, which could then 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.

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Section 9 Fram Strait.

The downstream conditions of the inflow to the Arctic Ocean though Fram Strait and the upstream conditions for the return flows to the strait are shown 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 the water returning to Fram Strait in the northern Nansen Basin, over the Gakkel Ridge and in the Amundsen Basin dominate.

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

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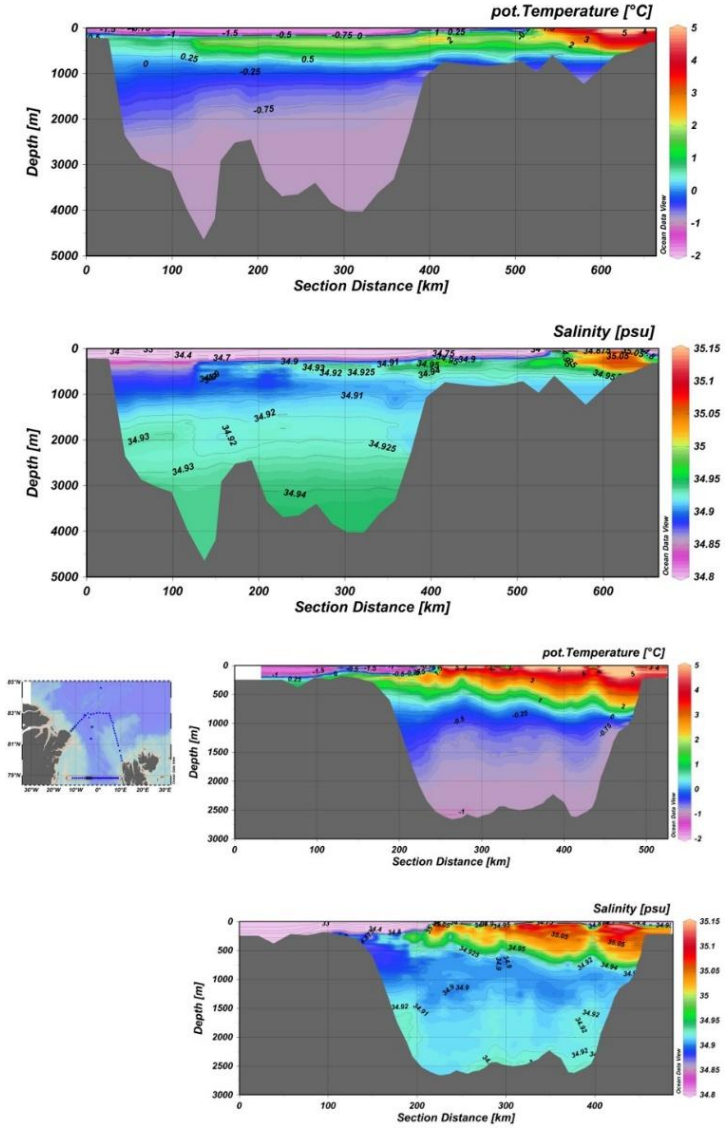

Figure 11. Two potential temperature and salinity sections, one taken from Svalbard along the Nansen Ridge and over the Yermak plateau into the Nansen Basin and the westward to Greenland abnd the second along the sill in Fram Strait showing the exchnges between t5he Arctic Ocean and the Nordic Seas (sections taken by *FS Polarstern* 2004).

On the southern section, taken along 78°50'N in Fram Strait (Figure 11), there is an inflow of mainly Atlantic water in the West Spitsbergen Current but also of intermediate and deep

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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 to 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, occasionally contributions from the Pacific inflows, and halocline water created from ice melted 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 also forms 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 1800m depth, which ultimately derives from the Amerasian 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 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

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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 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 that suggested by Nøst and Isachsen (2003), but also baroclinic transports of mainly less dense upper layer waters along the coastal boundaries (Werenskiold, 1935), 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 a westward recirculation in the strait.

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The presence of eddies 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 an opportunity 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 10^6 m^3s^{-1}, carrying a net salt export S of 13.973 kg^6s^{-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.

The northward and southward volume transports in Fram Strait are not balanced. This implies that 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.

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Computing the geostrophic transport through 16 hydrographic section 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 outflows 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 a mean temperature around -0.5^{o}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, $T_{ref} = 0.7^{o}$C and 25 TW obtained from the sections were taken to represent the transports and used to determined 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^{o}C enter the deeper layers and leave the Arctic Ocean through Fram Strait, this cold water transports $c\times(T_{ref} - (-0,5))\times1.2\times10^{9} = 5.8$ TW into the Arctic Ocean. $c = 4000$ Jkg^{-1}K^{-1} is the heat capacity of sea water. The remaining part of the net outflow of 1.6 Sv is less saline and at the freezing point and exits as upper layer water transporting $c\times(T_{ref} - (-1,8))\times1.6\times10^{9} = 16$ 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 almost all 25 TW of the

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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 $\phi = 2\alpha L(c\beta S_A)^{-1}$ (see section 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 made 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 indicates 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 Sv 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 mean 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 estimates of the Arctic Ocean heat and freshwater balances (e.g., Rudels et al., 2015). It appears that the heat and the effects of the heat carried by the Atlantic water into and around the Arctic Ocean are less than commonly expected. The existence of strong warm inflow pulses of Atlantic water could perhaps alter this situation and

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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.

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Section 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 homogenous water column southeast of Jan Mayen, and Amundsen's observations from *Gjøa* in the Greenland Sea in 1901 showed a cold homogenous 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 to convection studies 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

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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 homogenous 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). This idea was further developed 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

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the water column overturns more readily (Killworth, 1979, 1983). 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, 1967) no large positive anomaly could accumulate 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 startsZX (Rudels. 1986a). Since it is cold water that convects, it is possible that the thermobaric effect, arising because cold water is

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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 minutes and seven samples were collected. To observe the buildup and ejection of the saline top layer a 25m long drifting tube with temperature and conductivity sensors at 2.5m and 22.5m 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 22.5m. This perhaps could 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 figure 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 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.

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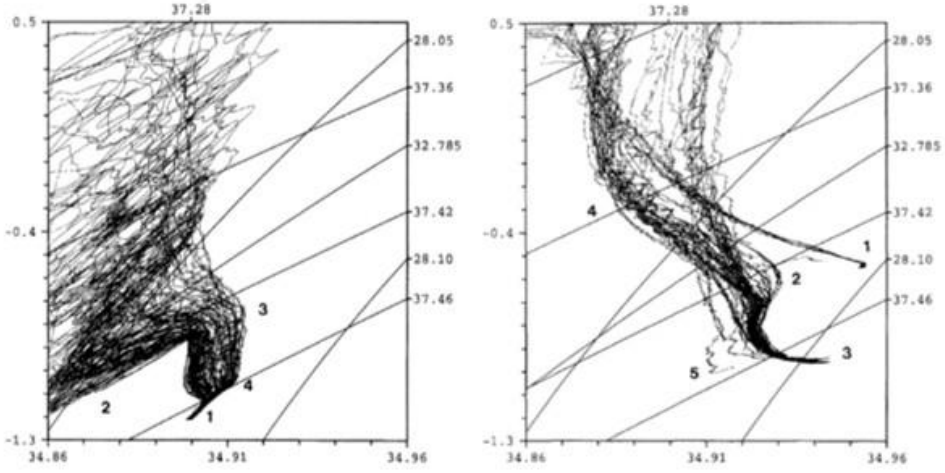

Figure 12. Left panel blow-up of θS curves from stations taken by *FS Valdivia* in 1993 in the Greenland Sea. The numbers indicate (1) Greenland Sea bottom water, (2) locally formed Arctic intermediate water, (3) intermediate temperature and salinity maxima, (4) deep salinity maximum. Right panel blow-up 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 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 1700m (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.

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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 supplying 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).

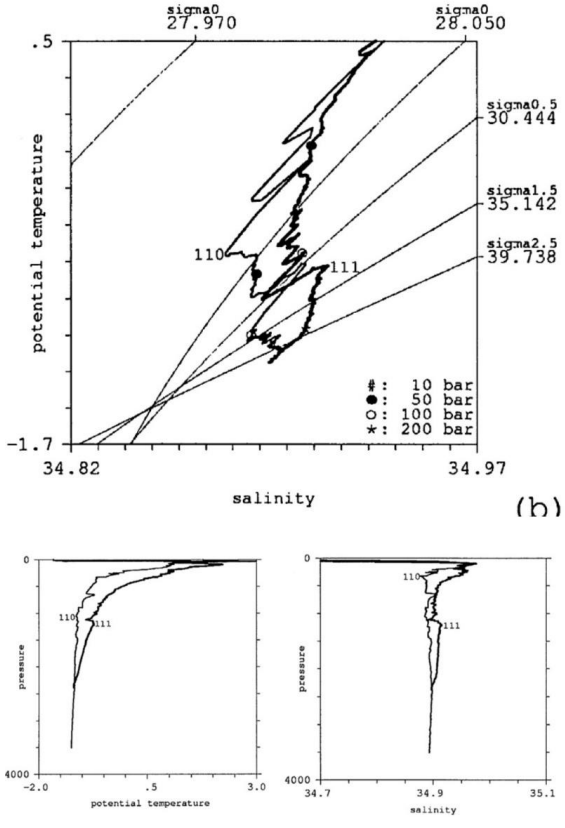

Figure 13. Potential temperature and salinity profiles and θS curves obtained in 1988 from *RV Lance* from stations over the Greenland continental slope. The stations show an isopycnal exchange of eddies between the slope and the central Greenland Sea (adapted from Rudels et al., 1999b).

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

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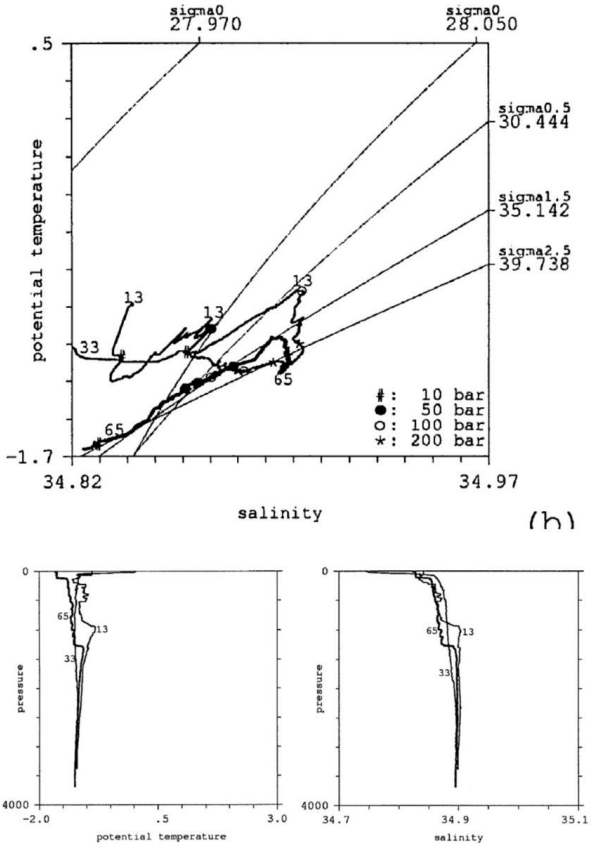
1596 density field in the central Greenland Sea, shifting the water from the rim upwards and that from the
1597 center to deeper levels at the rim. It is thus likely that the saline Arctic Ocean deep waters that
1598 penetrate to the inner part of the Greenland Sea become diluted by the convection taking place
1599 there largely removing their temperature and salinity anomalies.
1600



1601

1602 Figure 14. Potential temperature and salinity profiles and θS curves obtained by *FS Valdivia* in 1988
1603 (65 and 33) and 1993 (13) in the central Greenland Sea, showing the evolution of the convection in
1604 during 1988 and the shallower convection observed in 1993 (adapted from Rudels et al-., 1999b).
1605

1606 Figure 14 displays the potential temperature and salinity profiles and the θS curves of
1607 two stations taken in the center of the Greenland Sea in 1988, station 65 in February and station 33
1608 in June and a third station, 13, taken in 1993, all from *FS Valdivia*. The February station shows

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convection deeper than 1000m 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 indicate that convection has continued and reached deeper than 2000m. The salinity in the upper 1000m 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 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 is then increased sufficiently for it to convect into the second 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 by the lateral advection of Eurasian Basin deep water to the center of the gyre and the vertical mixing into the bottom water. The obtained mixing coefficient was surprisingly high, $\kappa_v \approx 3.7$ 10^{-3} m^2 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

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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.

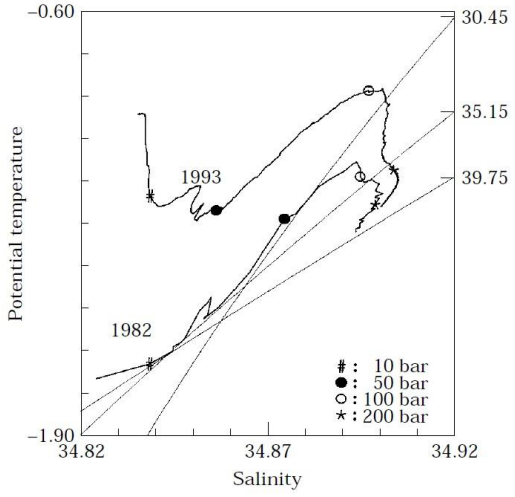

Figure 15. θS curves from two stations taken 11 years apart from *RV Hudson*, 1982 and *FS Valdivia* 1993.

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 no longer was 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 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

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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 vertical 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 m 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 only spread little vertically but

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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 created denser water at the surface that convected into the deep as entraining plumes that passed through the tracer cloud and redistributed the tracer to deeper levels. By mass continuity a compensating upward flow through the tracer cloud then brought 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 dominates and the Greenland Sea deep water is renewed by advection. If the Arctic Ocean source also have weakened, it has not yet been detected in the Greenland Sea because its larger reservoir in the Arctic Ocean. The cause behind these changes is likely the warming climate

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that no longer creates water dense enough to ventilate the deep Greenland Sea by open ocean convection.

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Section 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 2200m 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 indicated 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 Wywille Thomson first observed the cold water entering the Northeast Atlantic from the north.

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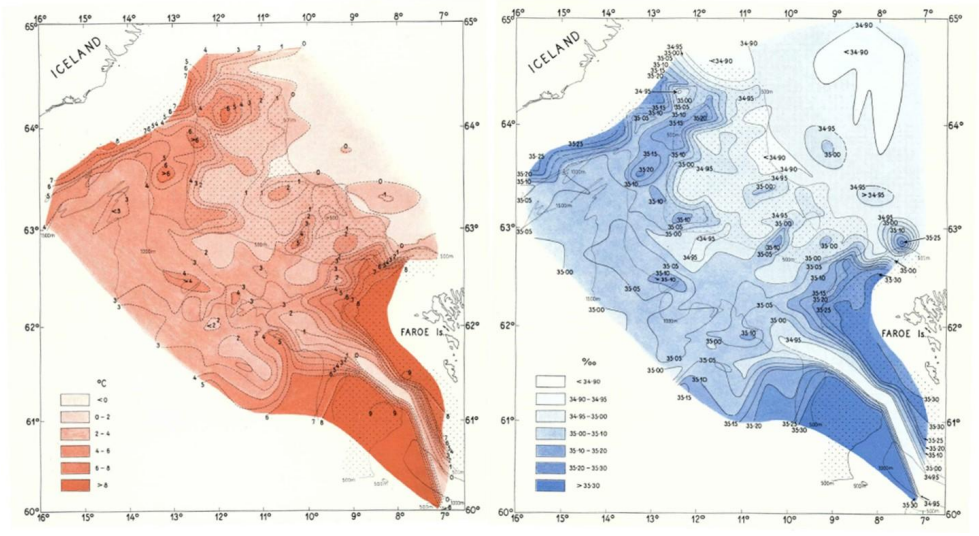

Figure 16. The bottom temperature (left) and salinity (right) distributions observed between Iceland and the Faroes during ICES's Overflow-60 experiment. Especially the cold and less saline water flowing through the Faroe Bank Channel is conspicuous compared to the cold water spilling over the Iceland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

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 Iceland 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 with a drifting depth of 600m in the Iceland Sea 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).

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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°C and 3°C and a salinity around 34.925.

The western overflow in Denmark Strait is more complicated. It is not a deep, almost stagnate 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 (1700m) 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 600m isobath on the northern Iceland shelf westward to Denmark Strait (Jónson and Waldimarson, 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 640m, 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., 1999).

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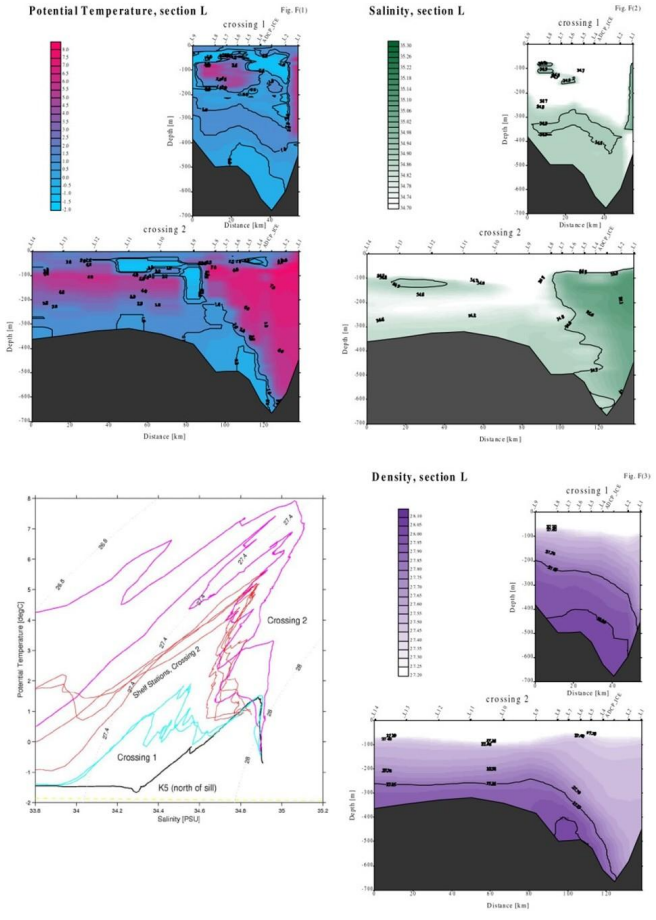

Figure 17. Potential temperature, salinity and potential density sections from two crossings between Greenland and Iceland along the sill in Denmark Strait taken one week apart. The θS diagram shows the θS curves from the two crossings, the first crossing in cyan and the second crossing in red and magenta.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

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 towards its unknown source area and it is observed along the entire northern

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shelf break (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 (Segtman 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 suppling 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, 1996b) did not find any strong seasonality of 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, 750m to 1500m, than the about 500m 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 600m 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; Rudels, 2021).

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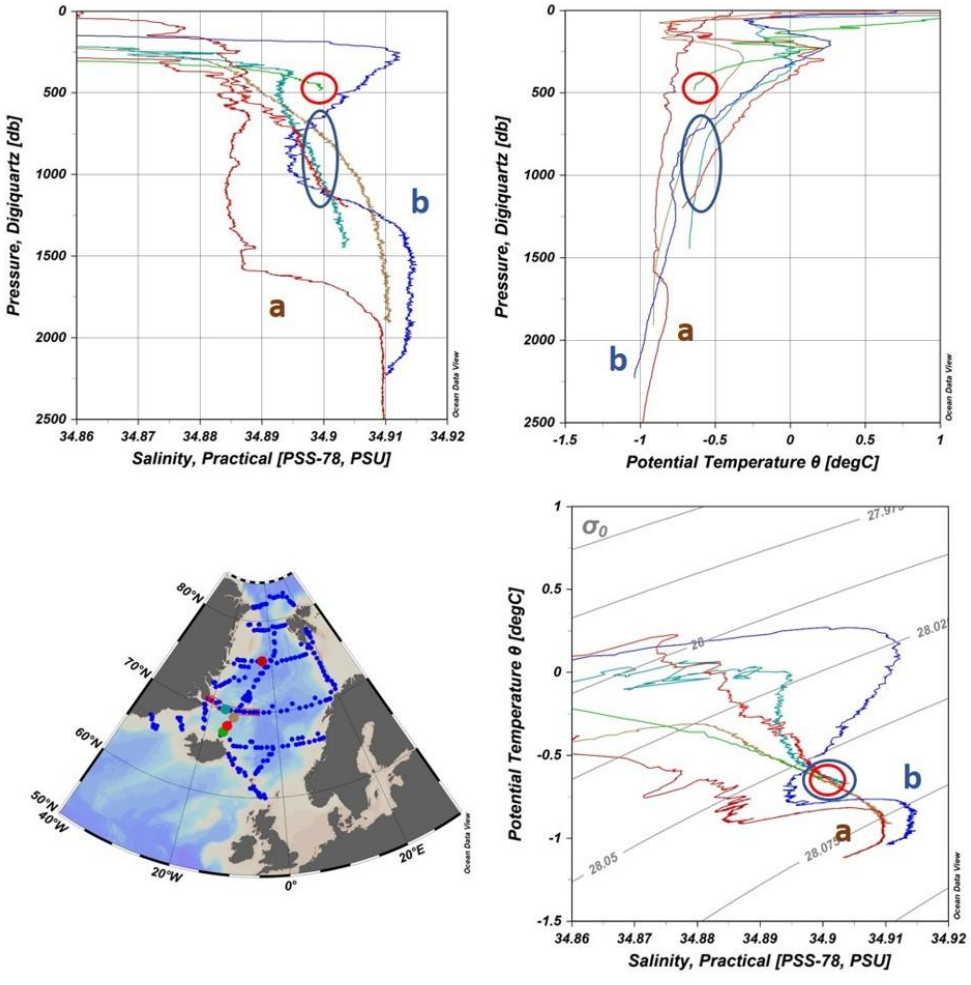

Figure 18. Profiles of potential temperature and salinity and θS curves from stations in the Iceland Sea and the Greenland Sea. The green station (red circles) found at 480 m depth at the Iceland slope has similar θS properties as the densest overflow and also properties akin to those observed below 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 from *RV Knorr* 2002 (Figure from Rudels, 2021).

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

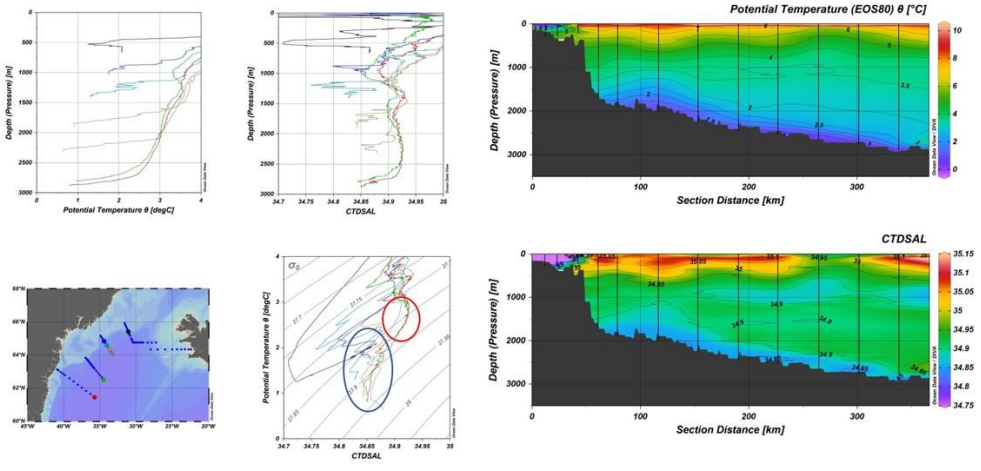

Figure 19. Potential temperature and salinity profiles and θS curves from the Greenland slope south of Denmark Strait. The sections show the overflow as a cold, less saline bottom layer, the warm Irminger Current in the upper layer and the presence of the more saline and warmer Iceland-Scotland overflow 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 (data from *RV Alexander von Humboldt* 2004) (Figure from Rudels, 2021).

The overflow plume in Denmark Strait is stratified. It comprises a cold, <- 0.5°C, bottom layer with salinity ≈ 34.9 below a temperature and a salinity maximum of 2°C and 34.91

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respectively, and it is capped by a salinity minimum, a low salinity lid (Figure 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 3000m in the Irminger Sea, and the low salinity 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 takes place already at the sill in Denmark Strait. There the deepest and densest part of the overflow becomes in almost direct contact with the warm and saline Irminger Current water, and the observed increase in plume temperature could result from a mixing ratio of 1 to 4, significantly 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 2000m 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 descend 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.

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 appears to 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., 2001). It flows as a separate core above the deep overflow plume towards Cape Farewell and eventually the density surfaces flatten and the spill jet disappear (Pickart et al., 2001). As the Denmark Strait plume descends it sinks

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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.

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 contributes, as estimated by the OSNAP west array, but little, 2.1 ± 0.3 Sv, to the overturning circulation. By contrast the transformation to denser water measured by the OSNAP east array was 15.6 ± 0.8 Sv, almost an order of magnitude larger. Since some recirculation takes place, the measured total overturning is somewhat less, 14.9 ± 0.9 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.

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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 rather forms MODE waters 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 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 leads to a cooling of the northeast Atlantic. This might break the warming trend in the Atlantic water that has prevailed almost two decades from the mid-1990s.

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Section 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, but 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, which 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).

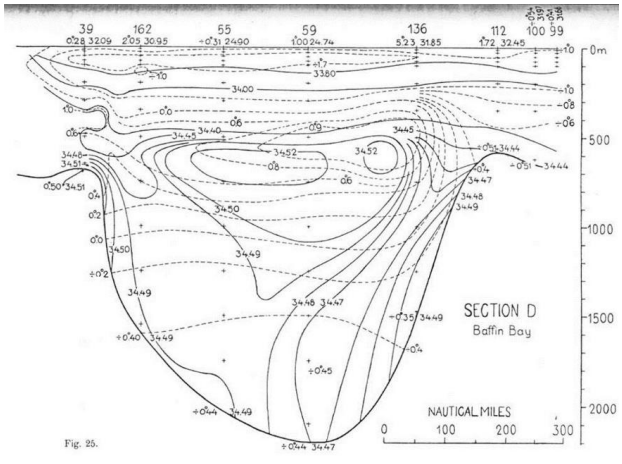

Figure 20. Temperature (dotted lines) and salinity (full lines) sections taken by *Godthaap* 1928 in Baffin Bay from Davis Strait (left) to Smith Sound (right), showing the cold upper layer, the intermediate temperature and salinity maxima and the cold, slightly less saline bottom water (from Riis-Karstensen,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

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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 been 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.

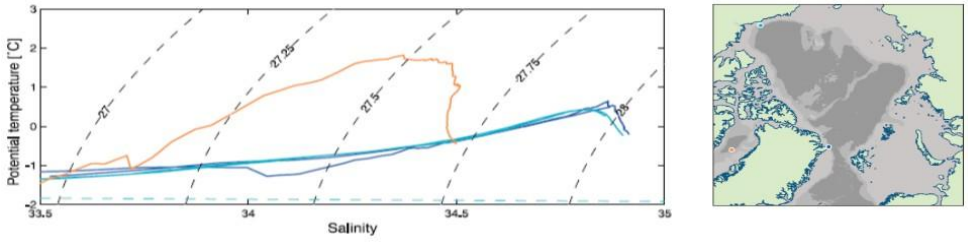

Figure 21. TS curves from Baffin Bay (yellow) and from north of Canada (cyan) and north of Greenland (blue), showing that the Baffin Bay bottom water could derive from the Arctic Ocean (from Rudels et al., 2004).

The 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

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much, it will become 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, 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 et al., 2008) 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 for 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).

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 Werenskiold (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.

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

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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 $g\beta(S_2 - S_1)$.

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 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. This gives:

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

The salinity of the upper layer becomes $S_1 = FH_1/(M_2 + F)$, where F is the freshwater input. The freshwater fraction in the upper layer, m, can then be written as:

$$m = \frac{(S_2 - S_1)}{S_2} = \frac{FH_1}{(M_2 + F)} = \left(\frac{2fF}{3g\beta S_2}\right)^{\frac{1}{2}} \quad (6)$$

And the total outflow becomes:

$$(M_2 + F) = \left(\frac{3g\beta F}{2f}\right)^{\frac{1}{2}} H_1 \quad (7)$$

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

$$Ro = \frac{[g\beta(S_2 - S_1)H_1]^{\frac{1}{2}}}{f} = \left(\frac{2g\beta S_2 F}{3f^3}\right)^{\frac{1}{4}} \quad (8)$$

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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 ms^{-2}, f = $1.4\times10^{-4}s^{-1}$, $\beta = 8\times10^{-4}$, F = 0.20×10^{6} m^3s^{-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 10×10^{12} m^2 and with a freshwater input 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 almost 50% by the 1990s (Rothrock et al., 1997), and if the ice export $F_i = 0.9$ Sv, remains constant, its residence time 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.

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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 density of the Baffin Bay upper layer determine the geostrophic transport out of Baffin Bay, which should balance the 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 (Figure 22).

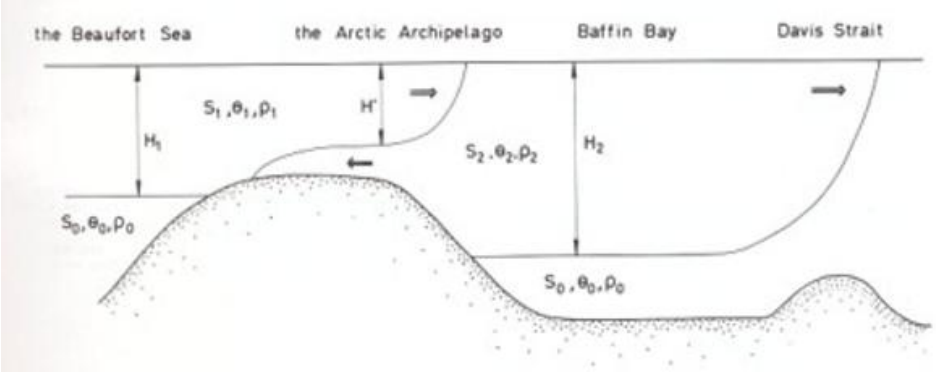

Figure 22. Idealized three reservoir system representing the Beaufort Sea, the Canadian Arctic Archipelago, Baffin Bay, Davis Strait and Labrador Sea system (from Rudels, 1986b).

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 (Rudels, 1986b). The transports decrease with increasing salinity of the Polar water, since the density difference between the layers then is reduced.

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However, the characteristics of the Baffin Bay upper layer are not due only 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 one 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, 2007), 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

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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, now 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 inflow recirculates in Baffin Bay and then exits through Davis strait together with, but above, the outflow of the deep waters from the Arctic Ocean (Fig. 23).

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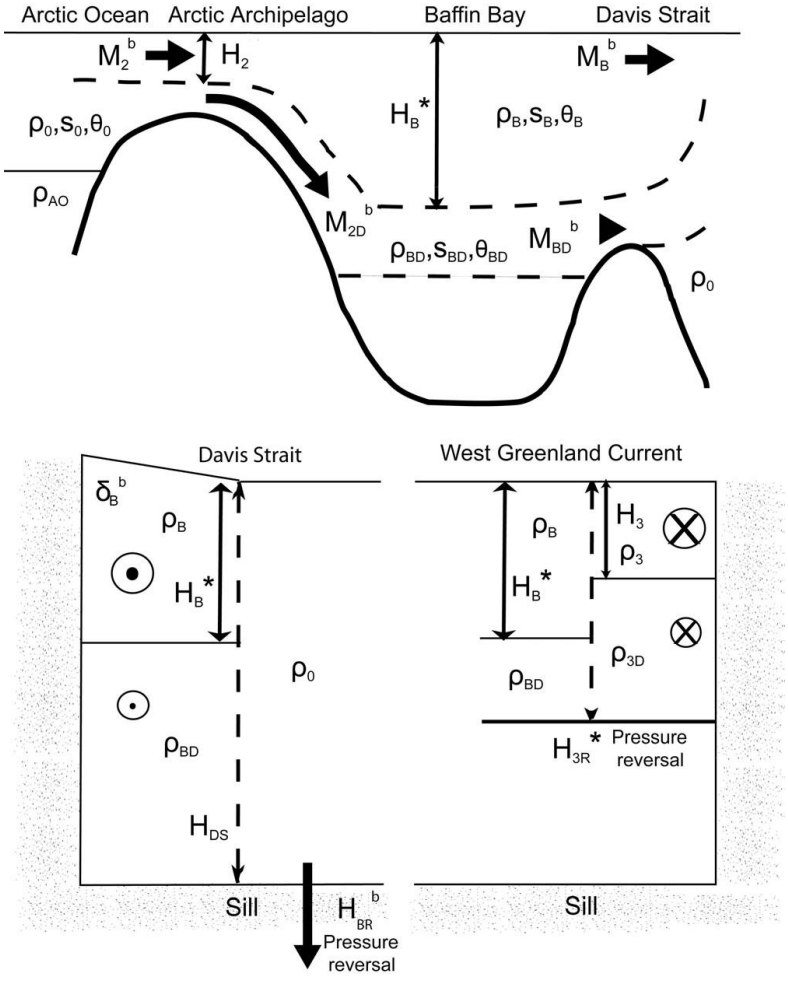

Figure 23. Drawing showing the inflow of Polar surface water M_2^b and denser water M_{2D}^b through Nares Strait and the outflow of upper layer water M_B^b and intermediate water M_{BD}^b through Davis Strait for the barotropic case. The depth of the upper layer H_3 and the depth of the pressure reversal H_{3R}^* in the West Greenland Current. The depth of the H_B^* and the sill depth H_{DS} in Davís Strait. The depth of the pressure reversal in the Davis Strait outflow H_{BR}^b is located below the sill depth. δ_B^b indicates the barotropic sea level difference between Baffin Bay and the Labrador Sea δ_3 is the density of the upper layer and δ_{3D} is the density of the intermediate layer in the West Greenland Current, δ_{BD} is the density of the intermediate layer in Baffin Bay and ρ_o is the density of the Labrador Sea water and of the deep inflow through Nares Strait. ρ_{AO} is the density of the deeper layer in the Arctic Ocean.

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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 upper layer outflow of 0.77 Sv. 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).

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Section 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 as warm, saline Atlantic inflows crossing the Greenland-Scotland Ridge. Only the Pacific inflow is 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 supplies a net input of salt to the Atlantic Ocean that crosses the equator and is exported to 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.

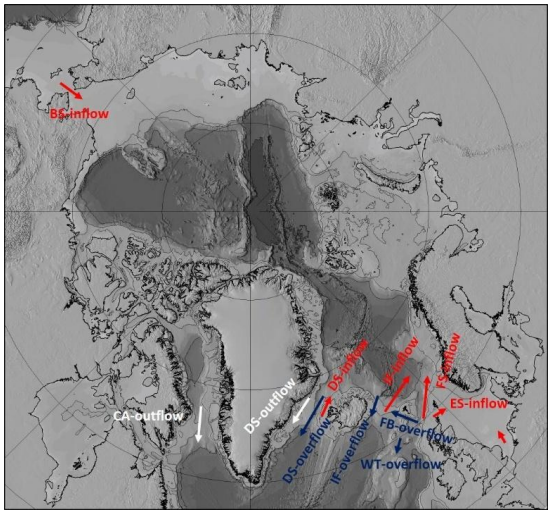

Figure 24: Map showing the volume balance of the Arctic Mediterranean Sea. BS: Bering Strait inflow, DS: Denmark Strait inflow, IF: Iceland-Faroe inflow, FS: Faroe-Shetland inflow, ES: European shelf inflow, CA: Canadian Arctic Archipelago outflow, DS: Denmark Strait outflow, DS: Denmark Strait overflow, IF: Iceland Faroe overflow, FB: Faroe Bank overflow, WT: Wyville Thomson overflow (adapted from Østerhus et al., 2019).

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Østerhus et al. (2019) compiled the measured volume transports through all different control sections to find if the volume fluxes balance (Figure 24 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 one 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.

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

Branch and abbreviation	Period yyyy/mm-yyyy/mm	months	Gaps Months	Transport Sv	Sd(Sv)
Inflows					
Denmarks Strait, DS-inflow	1994/10-2015/12	250	5	0.9 ± 0.1	0.3
Iceland-Faroe, IF-inflow	1993/01-2015/12	276	0	3.8 ± 0.5	0.6
Faroe-Shetland, FS inflow	1993/01-2015/12	276	0	2.7 ± 0.5	1.1
European shelf, ES-inflow	1993/01-2015/12	276	0	0.6 ± 0.2	0.3
Bering Strait, BS-inflow	1997/08-2013/12	197	0	0.9 ± 0.1	0.4
Overflows					
Denmark Strait, DS-overflow	1996/05-2015/12	218	18	3.2 ± 0.5	0.4
Iceland-Faroe, IF-overflow	NA			0.4 ± 0.3	
Faroe Bank, FB-overflow	1995/12-2015/12	206	35	2.0 ± 0.3	0.3
Wyville Thomson, WT-overflow	2006/05-2013/05	61	24	0.2 ± 0.1	0.1
Surface outflows					
Can.A.Arch., CA-outflow	2004/100-2010/09	72	0	1.7 ± 0.2	0.7
Denmark Strat, DS-outflow	2011/09-2012/07	11	0	2.0 ± 0.5	0.5
Runoff and precipitation					
Freshwater input, freshwater	NA			0.2	

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

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Tabell 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
Freshwater reservoirs (km^3)		
Liquid freshwater	93000	101000
Beaufort Gyre	18500	23500
Average sea ice cover	17800	14300
Total freshwater volume	110800	115300
Freshwater fluxes (km^3a^{-1}/mSv)		
Runoff	3900 ± 390 / 123.57 ± 12.36	4200 ± 420 / 133.08 ± 13.31
Bering Strait (liquid)	2400 ± 300 / 76,045 ± 9.51	2500 ± 100 / 79.21 ± 3.17
Bering Strait (sea ice)	140 ± 40 / 4.44 ± 1.27	140 ± 40 / 4.44 ± 1.27
P – E	2000 ± 200 / 63.37 ± 6.34	2200 ± 220 / 69.71 ± 6.97
Greenland flux	330 ± 20 / 10.46 ± 0.63	370 ± 25 / 11.72 ± 0.79
Davis Strait (liquid)	−3200 ± 320 / −101.39 ± 10.14	−2900 ± 190 / −91.89 ± 6.02
Davis Strait (sea ice)	−160 ± ?/ −5.07 ± ?	−320 ± 45 / −10.14 ± 1.43
Fram Strait (liquid)	−2700 ± 530 / −85.55 ± 16.79	−2800 ± 420 / −88.72 ± 13.31
Fram Strait (sea ice)	−2300 ± 340 / −72.88 ± 10.77	−1900 ± 280 / −60.20 ± 8.87
Barents Sea Opening	−90 ± 90 / −2.85 ± 2.85	−90 ± 90 / −2.85 ± 2.85
Fury and Hecla straits	−200 ± ? / −6.34 ± +	−200 ± ? / −6.34 ± ?
Total fluxes (km^3a^{-1}/mSv)		
Inflow sources	8800 ± 530 / 278.83 ± 16.79	9400 ± 490 / 297.85 ± 15.53
Outflow sinks	−8700 ± 700 / −275.67 ± 22.18	−8250 ± 550 / −261.41 ± 17.43
Residual	100 ± 900 / 3.17 ± 28.52	1200 ± 730 / 38.02 ± 23.13

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). The overflow water sinks into the deep and eventually joins the southward moving deep western boundary, but the outflow 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

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anomaly could be traced around the subpolar gyre and the 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 (Figure 25). In the late 1980s to the mid-1990s the North Atlantic Oscillation (NAO) index (the 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 increased 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 (Fig. 25).

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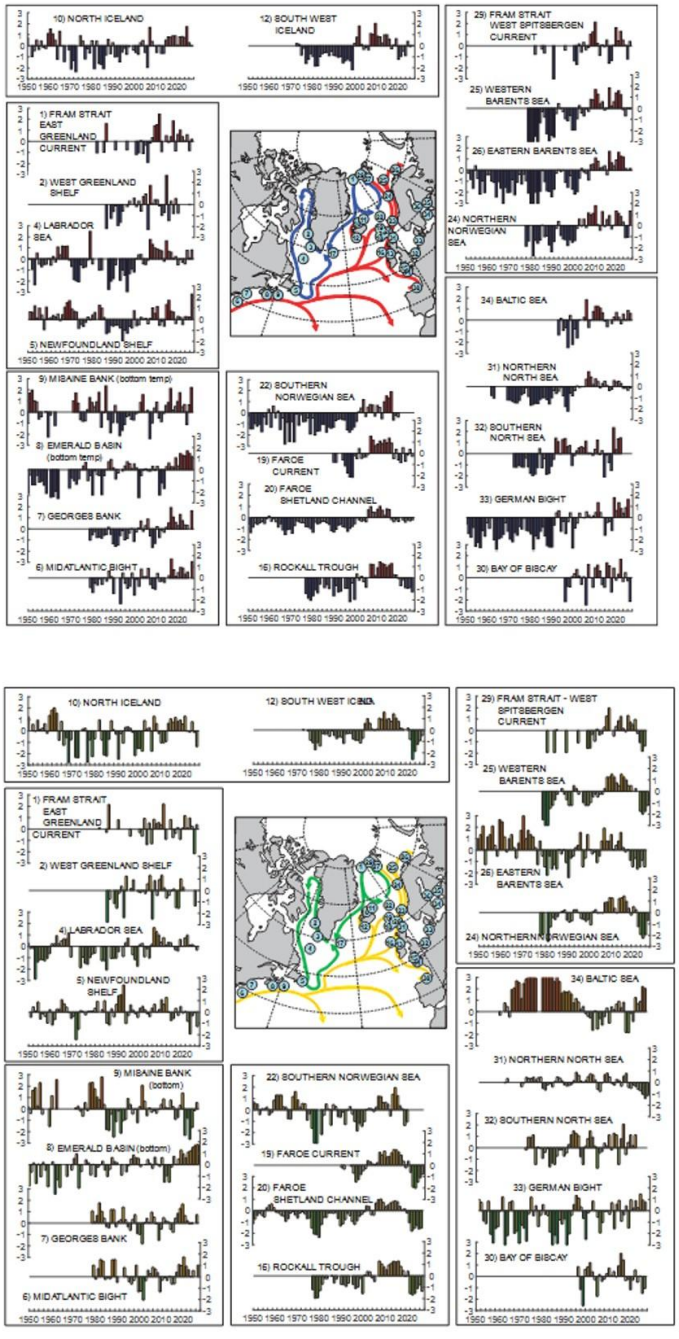

2199

2200 Figure 25. Positions of the ICES Report of Ocean Climate (IROC) standard surface temperature
2201 anomaly and salinity anomaly time series, and some selected temperature and salinity time series.
2202 The anomalies are from the long term mean temperature and salinity at the location (from González-
2203 Pola et al., 2023).

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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 ^{o}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 (Fig. 25).

In 2012 a strong freshening of the subpolar gyre was observed, the strongest on 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 (Fig. 25). 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

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and sinks into the deep. Moreover, it also appears to require a sill that allows the convection to build up a reservoir of dense water and also 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 relative 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, as It has been assumed before.

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