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

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22 Abstract

23 The Arctic Mediterranean is rapidly changing, a statement that is often made, but the follow-up
24 statements: changing from what and towards what are often omitted. Hence its role in the global
25 climate system, particularly regarding the Atlantic Meridional Overturning Circulations remains
26 poorly constrained. This review of the oceanography of the Arctic Mediterranean Sea develops an
27 unified perspective of how interacting components of the system evolve in space and time, and the
28 processes that determine their evolution. To set the stage a succinct overview of the geographic
29 setting and early explorations is given. We then follow the pathways of the principal water masses to
30 describe inflows, inter-basin circulations, water mass transformations, and outflows of heat and salt
31 to the bordering subpolar gyre and the global ocean. The fundamental connection to the global
32 ocean, the Atlantic water, is traced along its route into and through the Arctic Mediterranean. Its
33 transformations, driven by cooling, and by the freezing and melting of sea ice, lead to the creation of
34 both denser and less dense waters that form and maintain the water column structures within the
35 Arctic Mediterranean; the so-called double estuary. A second advective component to the Arctic, the
36 low salinity water carried by the Pacific inflow, is concentrated into the Amerasian Basin and acts
37 there to further isolate the denser waters derived from the Atlantic inflow. Hence, the waters return
38 to the North Atlantic either as dense overflows or buoyant outflows. The water masses within
39 diverse regions of the Arctic Mediterranean have changed over the past few decades, and this
40 influences their exchanges with the world ocean east and west of Greenland. That the water mass
41 transformations in the Arctic Mediterranean take place beyond a ridge system that allows for the
42 build-up of significant density differences that, through entrainment, can increase the impact of the
43 Arctic Mediterranean on global overturning circulation.

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47 Summary

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

54

55 Section 1 Introduction

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

66 Section 3 describes how our knowledge of the Arctic has evolved during the 500 years
67 that have elapsed since the exploration of the high northern latitudes started. The section also
68 reviews how the knowledge of the circulation and the water mass (trans)formations taking place in
69 the Arctic Mediterranean, ~~and how our understanding has gradually~~ evolved during the about 150
70 years that oceanographic observations and studies have been conducted in the Arctic
71 Mediterranean. It also discusses the early views of the deep circulation in the world ocean and finally
72 brings up the establishing of the Committee for the Exploration of the Sea (ICES) in 1899 and
73 discusses its importance for the improvement of measurement techniques and theoretical
74 understanding of the ocean circulation.

75 Section 4 examines the different inflows to the Arctic Mediterranean, the Pacific
76 inflow through Bering Strait and the North Atlantic circulation from the subtropical gyre to the
77 subpolar gyre and into the Nordic Seas. The next sections concentrate on specific features, areas and
78 processes that are presently active and may become more (or less) important in a future warmer
79 climate. Section 5 focusses on sea ice, its physical properties and how its growth and melting
80 influence the underlying water column. Section 6 describes the two inflows from the Nordic Seas to

81 the Arctic Ocean, through Fram Strait and over the Barents Sea, and the interactions (freezing and
82 melting) between Atlantic water and sea ice. Section 7 concentrates on the transformations of the
83 Atlantic water as it moves eastward along the two inflow paths and on the subsequent meeting and
84 remerging of the two inflow branches at the continental slope north of the Kara Sea, Severnaya
85 Zemlya and the Laptev Sea. In this section a wind driven model proposed by Nøst and Isachsen
86 (2003) is also presented.

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

99 In section 9 the water masses exchanged through Fram Strait are described from two
100 hydrographic sections, one north of the strait and one along the sill, both taken by *FS Polarstern* in
101 2004. Different water masses are identified and the transports of volume, freshwater and heat
102 through the strait are determined from geostrophy combined with inverse methods and a discussion
103 about the reasonableness of the obtained mass and freshwater transports is provided.

104 The Nordic Seas, especially the Greenland Sea, are characterized by weak stability and
105 are areas of deep convection and deep water renewal, and the convection in the Greenland Sea is
106 discussed in section 10. The water column in the Greenland Sea has exhibited large changes during

107 the last 50 years, indicating that the nature of the active convection processes has changed
108 (compare, for example Carmack and Aagaard, 1973 and Brakstad et al., 2019). This is perhaps the
109 area, where the effects of a warming climate on the ~~deep~~ water column have been the strongest.

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

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

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

130 The present review ~~is intended~~intends to give a comprehensive ~~discussion~~picture of
131 the Arctic Mediterranean Sea, describing the processes and the water mass transformations in the
132 different parts of the area and how these parts are connected by the circulation of the different

waters. The manuscript also highlights how our knowledge of the geography and our understanding of the physical oceanography of the Arctic Mediterranean have evolved over time. The focus is on observations and their interpretation. Processes are described and their possible importance discussed, occasionally by the use of conceptual models. For different aspects of and/or more detailed descriptions of the oceanographic processes active in the Arctic Mediterranean we refer the readers to the review articles by Coachman and Aagaard (1974), Timmermans and Marshall (2020), Rudels and Carmack (2022), Wassman et al., 2020, and the books by Lewis et al. (2000), Dickson et al. (2008), Mills (2009) and Rudels (2021) and Wassman (2024) and to the references given in these studies. The narrative follows two tracks. The first is historical and describes how our knowledge of the geography, the hydrography and the circulation has evolved through time. This approach continues to about 1980 and encompasses the knowledge provided by early summaries of the circulation in the Nordic Seas (Helland-Hansen and Nansen, 1909), the connections between the North Atlantic and the Nordic Seas (Helland-Hansen, 1912; Nansen, 1912; Nansen, 1913) and the early knowledge of the water masses and circulation in the Arctic Ocean (Nansen, 1902; Nansen, 1906; Timofeyev, 1960; Coachman and Aagaard, 1974).

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

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

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

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

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

177 The reviews most akin to the present one are written by Rudels (2019), Timmermans
178 and Marshall (2020), and Rudels and Carmack (2022). Rudels (2019) gives a concentrated description
179 of the Arctic Ocean circulation and water mass transformations, while Timmermans and Marshall
180 (2020) describe the dynamically distinct Arctic circulation: wind forcing, buoyancy forcing,
181 topographic steering, eddies, and boundary currents. They emphasize how freshwater storage, sea-
182 ice loss, and changing Atlantic or Pacific inflows may reshape stratification and circulation. Rudels
183 and Carmack (2022) discuss the Atlantic and Pacific inflow branches, halocline formation, shelf

184 contributions, and deep-water renewal, and stresses the controlling role of ridges, shelves, and
185 narrow gateways in partitioning water masses among the Arctic basins.

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

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

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

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

206 This review is, admittedly, focused on Western science, as based on the scientific
207 method, and gives little recognition to indigenous contributions to exploration and ecological
208 knowledge. There are two main reasons for this. First, it is beyond the scope of this review, and the
209 authors lack the necessary expertise to do this complex subject justice. Second, the focus of this

review is on the circulation and transformations of waters in the deep and offshore Arctic
Mediterranean, and thus remote from the Arctic coastal regions where indigenous peoples travel,
hunt and live. Regardless, there is much to learn from indigenous thinking as our global society
addresses an uncertain future (cf. Berkes and Joly, 2001; Ford et al, 2010; Falardeau et al., 2019a, b;
Huntington, et al., 2021). Indigenous knowledge resides in the holder's authority and is experiential
and holistic and place-based (Gearheard, S., et al., 2006; Whiteman and Cooper, 2011; Fienup-Riordan
and Carmack, 2011). Indigenous Knowledge is not a single, uniform body of information, but is
learned, practiced, and transmitted differently among distinct Arctic Indigenous Peoples; for reviews,
see Eerkes-Medrano and Huntington (2021). Despite the above distinctions, there are key areas
where complementarity exists (cf. Carmack and Macdonald, 2008). Intimate connections and
dependence of snow and ice conditions can provide an early-warning system for climate and
ecological change; one clear example is in documenting local observations based intergenerational
knowledge and travel/hunting practice as records of environmental change (Krupnik and Jolly, 2002;
Carmack, 2021). That the two-ways of knowing are jointly beneficial was demonstrated by
comparison of remote sensing of sea ice and Inuit observations (Meier et al., 2006). Fienup-Riordan
and Carmack showed the complementarity of broad spatial records (wind-generated ocean swell) the
and fine-scale knowledge of ice quality and currents (safety of travel over ice in coastal locations).
We emphasize that the depth, breath and sophistication of Indigenous Knowledge of what is Arctic
equals that of Western science, and future avenues based on appropriate co-generation of
knowledge offer a rich field of inquiry.

231 Section 2 The North Polar Setting

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

242 The earth's rotation axis makes an angle of 23.5° to the ecliptic. This leads to large
243 seasonal variations in incoming solar radiation, especially north and south of the polar circles
244 ($>66.5^\circ$). In summer the incoming daily radiation at the top of the atmosphere is larger at the poles
245 than at the equator, (e.g. Fritz, 1951), while in winter the sun is below the horizon and the long, dark
246 polar night reigns. The contrast between the two polar regions is, however, large. In the south the
247 entire ice covered Antarctic continent is in direct contact with the circumpolar Southern Ocean, while
248 in the north there is a confined ocean, the Arctic Mediterranean Sea, with restricted communications
249 to the world ocean and with its northernmost part, the Arctic Ocean, partly covered by perennial sea
250 ice.

251 The atmospheric poleward heat transport is carried by three distinct circulation cells,
252 the direct Hadley cell where warm air rises at the equator, cools and then sinks back to the surface at
253 20° to 25° N(S). Some of the sinking air returns towards the equator, ~~creating~~forming the trade
254 winds, while the rest is transported poleward by the westerlies in the eddy dominated Ferrell cell.
255 Over the polar areas, in the direct Polar cell, the air is cooled and returns equatorward forcing the
256 warmer air of the Ferrel cell to rise.

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

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

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

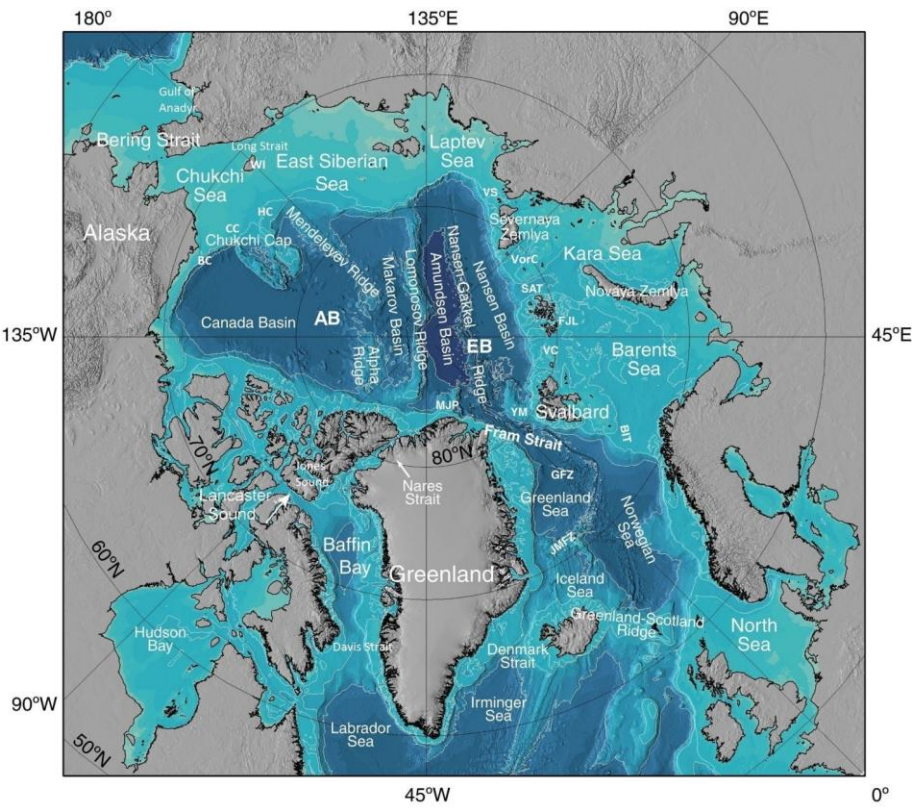
278 The surface air temperature in the Arctic is above the melting temperature of sea
279 water in summer but much below in winter. This leads to phase changes. Sea ice is formed in winter
280 and melts again in summer. The latent heat released by freezing becomes a local heat source that
281 somewhat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

283 thermostat that keeps the ~~water~~-surface water temperature close to the melting temperature as long
284 as ice is present. Ice formed and melted does not contribute any net heat to the Arctic Ocean, but
285 the melting in summer stabilizes the water column and isolates the deeper layers from the surface
286 forcing in summer even more than in winter. The ice not melted in summer is eventually exported by
287 the ocean currents southward and becomes a heat source to the Arctic.

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

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

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312 Figure 1: Map of the Arctic Mediterranean Sea showing geographical and bathymetric features. The
313 bathymetry is from the IBCAO updated data base and the projection is Lambert Equal Area. The 200,
314 500, 2000 and 4000 m isobaths are shown. AB; Amerasian Basin, BIT; Bear Island Trough (Barents Sea
315 opening), BC; Barrow Canyon, CC; Central Channel, EB; Eurasian Basin, FJL; Franz Josef Land, GFZ;
316 Greenland Fracture Zone, HC; Herald Canyon, MJP; Morris Jessup Plateau, JMFZ; Jan Mayen Fracture
317 Zone, SAT; St. Anna Trough, YM; Yermak Plateau, VC; Victoria Channel, VorC; Voronin Canyon, VS;
318 Vilkitskij Strait, WI Wrangel Island (Adapted from Rudels 2019, figure 1).

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321 The deeper part of the Arctic Ocean consists of two major basins, the Amerasian Basin
322 and the Eurasian Basin, separated by the narrow Lomonosov Ridge with ~~an average~~ a mean sill depth
323 around 1600 m. The Amerasian Basin is divided by the Alpha and Mendeleev ridges into the large
324 Canada Basin and the smaller Makarov Basin, both with maximum depth of the abyssal plains around
325 4000 m. The Eurasian Basin also consists of two basins, the about 4000 m deep Nansen Basin
326 between the Eurasian continental slope and the Gakkel Ridge and the deeper, 4500 m, Amundsen
327 Basin extending from the Gakkel Ridge to the Lomonosov Ridge. The connection to the Nordic Seas
328 through Fram Strait is deep, sill depth 2600 m. The Greenland Sea comprises two basins, the
329 Greenland Basin with maximum depth about 3500 m and the northerly, smaller and shallower
330 Boreas Basin. The Norwegian Sea is separated into the Norwegian Basin and the Lofoten Basin both
331 about 3000 m deep, while the Iceland Sea is shallower, less than 2000 m deep.

332 Three deeper passages ~~across~~ through the Greenland-Scotland Ridge connect the
333 Nordic Seas to the North Atlantic, the 640 m deep Denmark Strait between Greenland and Iceland,
334 the 400 m to 500 m deep ridge between Iceland and the Faroes, and the channel between the Faroes
335 and Scotland, ending in the more than 800 m deep Faroe Bank Channel. West of Greenland the
336 channels from the Arctic Ocean to Baffin Bay are narrow and shallow. Nares Strait has a sill depth of
337 220 m and Lancaster Sound and Jones Sound have sill depths between 100 m and 150 m. Baffin Bay
338 is deep, about 2600 m, and the sill depth in Davis Strait to the Labrador Sea is 630 m. Bering Strait
339 between the North Pacific to the Arctic Ocean is narrow, 85 km, and shallow, sill depth 53 m.

340

341 Section 3 Exploration

342 Section 3.1 voyages of discovery

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

355 The efforts to find a northern route between the Atlantic and Pacific continued,
356 however, and the possibilities to find both a Northwest Passage through or north of North America
357 and a Northeast Passage north of the Eurasian continent were explored but with no success. The
358 expeditions were stopped by ice and many ended with the loss of ship and crew. It would take 300
359 years before both passages had been navigated by a single vessel, the Northeast Passage by *Vega*
360 1878-1880 and the Northwest Passage by *Gjøa* 1903-1905.

361 The French explorer Jacques Cartier entered the Gulf of St Lawrence and the St
362 Lawrence river in 1535, which proved to be an entrance for France to the interior of North America,
363 but found no passage to the Pacific. The English search eventually turned to higher ~~higher~~ latitudes.
364 In 1576 Martin Frobisher sailed west of Greenland and entered a passage that he thought might lead
365 to the Pacific, but it was just a bay, later named Frobisher Bay. John Davis made two voyages west of
366 Greenland and north of Frobisher Bay and on his return voyage he noticed a strait with strong

367 currents south of Frobisher Bay that could ~~be~~indicate a passage west. In 1610 Henry Hudson was
368 given the task to ~~examine~~explore this passage, but after sailing through the strait, later named after
369 him, and entering the large open water area autumn was approaching ~~and the~~. The ship became
370 trapped in the ice and the expedition had to overwinter. In spring when the ship came loose of the
371 ice there was a mutiny and Hudson, his son and seven members of the crew were put in a lifeboat to
372 face a certain death.

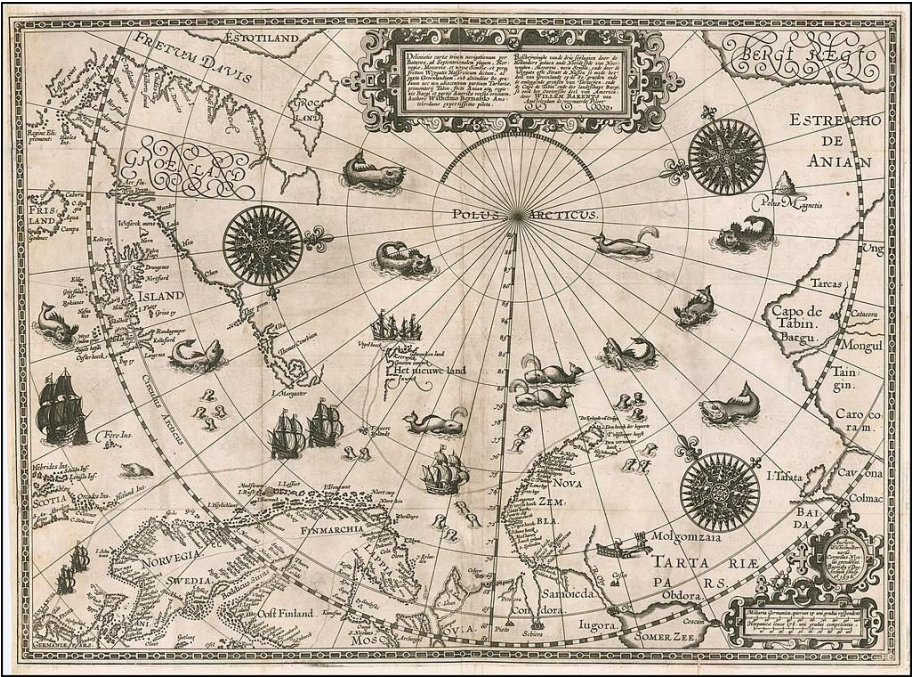
373 ~~When~~As the ship returned to London the crew presented Hudson's sailing notes and
374 was pardoned. Further expeditions showed, however, that Hudson had sailed into a bay,
375 appropriately now named Hudson Bay, not into a passage to the Pacific. In 1616 Robert Bylot, with
376 William Baffin as pilot instead sailed as far north as they could west of Greenland and discovered
377 three possible passages, Smith Sound, Jones Sound and Lancaster Sound but they judged that
378 ~~they~~these were not leading to the Pacific. Baffin's sailing notes were, unfortunately, never published
379 and the expedition was all but forgotten.

380 The English efforts to find the Northwest Passage eventually led to the establishing of
381 one early successful commercial venture in the new world, the Hudson Bay company, and the search
382 for the Northeast Passage resulted in contacts with Russian traders at the river Dvina and to the
383 formation of the Muscovy company. This put an end to the English efforts to navigate the Northeast
384 passage but the search continued, now mainly sponsored by Dutch merchants. On three voyages in
385 1594, 1595 and 1596 with Willem Barents as pilot the Dutch reached Novaya Zemlya but could not
386 penetrate farther east. The 1596 expedition took a more northerly route and sighted Bear Island and
387 Spitsbergen before it turned east and reached the northern cape of Novaya Zemlya. Here the ship
388 was trapped in the ice and the expedition members overwintered on land. The following spring they
389 tried to reach the Kola peninsula in a small boat, but Barents ~~had~~died already ~~died~~ at the beginning
390 of the sailing.

391 The discovery of the northern seas with their abundance of life, whales and walrus, was
392 found more rewarding than failed searches for an unknown Northeast Passage and an intensive

393 whaling industry was established, mainly by the Dutch, on Spitsbergen. Holland had now become
394 independent from Spain and Spain had given up its monopoly of the southern routes to Cathay. No
395 pressing need for the Dutch to find the Northeast Passage then remained (Figure 2).
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400 Figure 2: A map of the Arctic Ocean by Willem Barents drawn around 1598 showing the then known
401 features of surrounding lands and islands. The ships north of Svalbard probably indicate open water,
402 the Whalers' Bay (from Häkli, 1992).

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404

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

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

419 Bering had not completed the charting of the strait between Asia and North America.
420 This was done by James Cook on his last voyage in 1779, and a later expedition, 1791 to 1795,
421 commanded by George Vancouver, finally proved that the Strait of Anian did not exist. Any ship
422 sailing into the Arctic Ocean from the North Atlantic to the North Pacific would have to exit through
423 Bering Strait.

424 About the same time the possibility to sail directly across the Arctic Ocean to Bering
425 Strait was considered. One early, but failed, attempt to cross the Arctic Ocean had been made by
426 Henry Hudson ~~already~~ in 1608. In spite of the fact that all previous expeditions had been stopped by
427 ice, the existence of an ice free inner part of the Arctic Ocean was seriously discussed. Robert
428 Thorne, an English merchant, had already in 1527 suggested that the continuous daylight during
429 summer should store enough heat in the water column to prevent ice formation in winter. Another
430 idea, favored by the Dutch mapmaker Peter Plancius, was that an ocean in perpetual motion would

431 not freeze. The Russian scientist Michail Lomonosov suggested that ice can only form in connection
432 with land and river runoff and the interior of the Arctic Ocean might be free of ice. In 1772 the Swiss
433 geographer and librarian Samuel Engel noticed that sea ice contains little salt and proposed that sea
434 water could not freeze. These ideas led to a few failed attempts to sail to the Pacific via the North
435 Pole.

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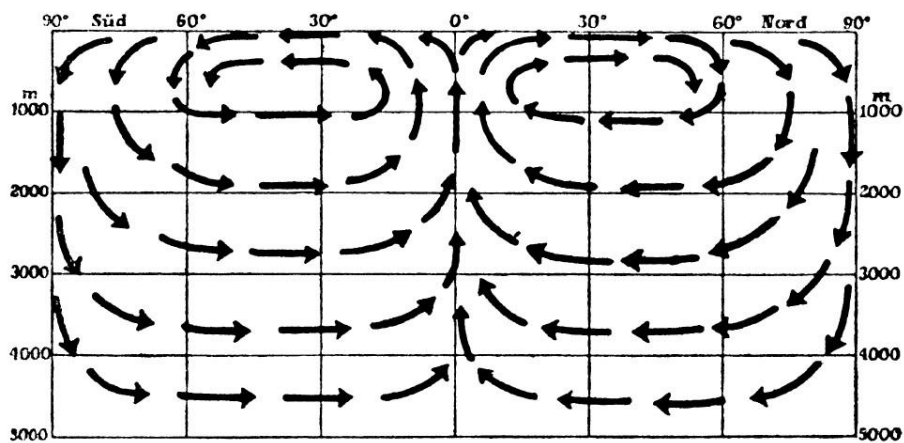
437 3.2 The deep circulation

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

447 In 1751 captain Henry Ellis on the vessel *Earl of Halifax* made temperature
448 measurements in the deep tropical Atlantic by collecting water from different depths in a bucket
449 equipped with a thermometer. He found that the temperature decreased to about 9°C at 1200 m but
450 then stayed constant, as the measurements continued down to 1600 m. These observations,
451 although reported in the Philosophical Transaction of the Royal Society, did not receive much
452 attention until Benjamin Thompson (Count Rumford) almost 50 years later proposed that the cold
453 deep water was advected from higher latitudes, where the winters were cold enough to create water
454 of the observed temperature. Thompson (1797) also pointed out that ocean water does not attain its
455 maximum density at 4°C, as in freshwater, but instead its density increases with decreasing
456 temperature until freezing.

457 This was demonstrated by Emil von Lenz, who made deep temperature measurements
458 using thermometers protected in self-closing buckets on Otto von Kotzebue's second
459 circumnavigation in 1823-1826, and observed temperatures well below 4°C. Von Lenz returned to
460 the study of the ocean deep circulation later in his life and proposed a thermally driven deep
461 circulation symmetric about the equator, the water sinking at high latitudes and rising towards the
462 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).

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

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

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

499 in the deep, are driven was far from solved, and its complexity hardly appreciated by either
500 Carpenter or Croll (Mills, 2009).

501

502 3.3 The Nordic Seas

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

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

523 Mohn assumed that wind drives the surface water in the same direction, but as the
524 water is moving on a rotating earth the current becomes deflected to the right (northern

hemisphere), leading to a change in sea level slope that eventually balances the Coriolis force and again makes the current parallel to the wind. The wind effect was added to the computed sea level distribution and found to be larger than and reinforcing the sea level slope due to the density distribution (Mohn, 1887).

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

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

3.4 Warm Currents and ice cover

Croll had emphasized the importance of the heat carried by the Gulf Stream for the European climate, but warm currents that enter the Arctic Ocean might also melt ice and lead to an open Arctic Ocean beyond the region influenced by continental runoff. One strong advocate for an ice-free inner Arctic Ocean was the geographer August Petermann, who examined the path of and the heat carried by the Gulf Stream and traced Gulf Stream water along the Siberian shelf as far as Bering Strait

551 (Petermann, 1865). He also suggested that the best route to enter the inner part of the Arctic Ocean
552 would be over the Barents Sea east of Svalbard.

553 Petermann was perhaps the most ardent believer in an open Arctic Ocean and of the
554 importance of warm ocean currents, but he was not the only one. Captain Silas Bent (1872) argued
555 that the pathways into the Arctic Ocean were in the wake of the warm currents, and Van Campen
556 (1877) tried to convince the Dutch authorities to revive the Dutch Arctic explorations, which had
557 been so intense in the 17th century, by sending expeditions following different warm currents into
558 the Arctic Ocean.

559 However, there were also contrary views. Maury (1855) argued that since the
560 southward flowing cold water was found at the surface and had low salinity, it was instead warm and
561 saline water from the south that rose from the deep and brought heat to the surface, which melted
562 the ice and kept the inner parts of the Arctic Ocean open. The low salinity mixture of sea ice and cold
563 water then flowed south in the East Greenland Current and the Labrador Current. This would then
564 ~~be~~lead to a circulation akin to that suggested by von Weitz (1755) (see below).

565 Efforts to reach the open Arctic Ocean were made at all passages where warm surface
566 water was observed to flow north. Elisha Kane (1853-1855), Charles Hall (1866-1871) and George
567 Nares (1874-1875) tried to enter the Arctic Ocean west of Greenland, Julius Payer and Karl
568 Weyprecht sailed 1872 with the ship *Admiral Tegetthoff* towards the Arctic Ocean east of Svalbard,
569 and George Washington de Long (1879-1881) entered through Bering Strait with *Jeanette* in 1879. All
570 these efforts failed and on several expeditions the ships were trapped in and eventually crushed by
571 the ice and the expedition members had to man the life boats. Some made it back, some perished.
572 The *Tegetthoff* expedition discovered Franz Josef Land but then the ship was caught in the ice and
573 crushed. The expedition members made it back in the life boats.

574 The *Jeanette* was trapped by the ice north of Bering Strait and drifted with the ice for
575 21 months in no obvious direction. After almost two years *Jeanette* broke up and vanished beneath
576 the ice. The expedition members were prepared and had put provisions, sledges and boats on the

577 ice, and they started the journey south towards Siberia. They reached open water in early August and
578 took to the boats, but a storm separated the three boats. One boat disappeared, one boat reached
579 the Lena river and the men were saved. De Long's boat made it to the Siberian coast, but they did
580 not meet anyone to help them finding food and all died by starvation. Three years later, in 1884,
581 objects from the *Jeanette* expedition were found by inuits on an ice floe close to Julianehåb on the
582 southwestern coast of Greenland, indicating that they had been carried by the ice from the wreckage
583 across the Arctic Ocean to the Nordic Seas and along the east coast of Greenland down to and
584 around Cape Farewell.

585

586 3.5 The drift of *Fram*

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

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

602 research vessel with laboratories to work on the observations and the samples collected during the
603 drift. *Fram* was in fact the first drifting scientific ice station.

604 *Fram* sailed from Oslo in June 1893 and in September she had reached the New
605 Siberian Islands, where she turned northward into the ice and was tied to an ice floe. The drift,
606 lasting until summer 1896, had begun. The first winter showed that *Fram* could withstand the ice
607 pressure. Meteorological and hydrographic observations were made regularly and depth
608 measurements indicated water depths close to 4000 m. The tacit assumption that the Arctic Ocean
609 was shallow was wrong, however, and the possibility of finding large land masses in the Arctic Ocean
610 was ruled out.

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

621 The knowledge gained by the drift of *Fram* did not, however, remove the sea ice and
622 the obstacle it was for further studies of the Arctic Ocean, and in the following forty years
623 oceanographic studies were mainly confined to the open areas and the marginal ice zone during the
624 summer season. Plans for a submarine expedition to the North Pole with the discarded submarine
625 *Nautilus* were made in 1931, but failure with the diving rudders prevented the submarine to dive.
626 The original plans then had to be abandoned and the observations were confined to the northern

627 part of Fram Strait and the area north of Svalbard (Sverdrup and Soule, 1933). In the late 1930s the
628 Soviet ice breaker *G. Sedov* was trapped in the ice and involuntarily repeated the drift of *Fram*.

629 A major step forward in the research of the interior of the Arctic Ocean occurred in
630 1937, when the Soviet Union established the first drifting ice station, NP-1, at the North pole. The
631 station drifted rapidly south and the team was picked up by Soviet ice breakers in February 1938 in
632 the Greenland Sea. After the Second World War new ice stations were launched, mainly by the Soviet
633 Union, but also by the United States and Canada. The Soviet Union also organized airborne
634 expeditions each spring covering the entire Arctic Ocean, where water sampling, ice thickness
635 measurements and meteorological observations were made. The observations from the ice stations
636 and from the airborne expeditions provided the knowledge of the ice drift, the ocean circulation and
637 the water mass distribution in the Arctic Ocean as it existed around 1980, before the beginning of the
638 ice breaker and the submarine expeditions and the revolution in remote sensing and in mooring
639 techniques that have taken place the last 40 years (see e.g., Rudels, 2015, 2021~~);~~)

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

650

651 3.6 ICES and technical and theoretical advances.

652 The drift of *Fram* was part of the tremendously active period in oceanography starting with the

653 *Challenger* expedition 1874-1876. Between 1899 and 1902 the International Committee for the
654 Exploration of the Sea (ICES) was established with its secretariat in Copenhagen and the ICES
655 laboratory in Kristiania (Oslo) with Nansen as director. One of the first task of ICES, ~~really~~actually the
656 reason for its founding, was to coordinate a five years observation program, where the different
657 member states should arrange oceanographic expeditions in all seasons in their adjacent waters,
658 thus covering much of the northern North Atlantic, the Norwegian Sea, the Barents Sea, the North
659 Sea and the Baltic.

660 As a response the Norwegian Fishery Board launched an extensive research program
661 in the Norwegian Sea between 1900 and 1904 with the research vessel *Michael Sars*. The report on
662 these expeditions, *The Norwegian Sea*, published by Bjørn Helland-Hansen and Fridtjof Nansen in
663 1909, described not only observations made within the program, but almost all available
664 observations and measurements were included in the report. *The Norwegian Sea* discussed all the
665 major water masses and currents and most of the major circulation features; the Atlantic water
666 entering from the south and flowing northward in the Norwegian Atlantic Current, the Norwegian
667 Coastal Current ultimately from the Baltic Sea and moving along the Norwegian Coast, the southward
668 flowing East Greenland Current and its branches the Jan Mayen Current and the East Iceland Current,
669 and the major cyclonic gyres in the different basins were all identified. Not only surface currents but
670 also the distribution and motions of the waters in the deeper layers were deduced and described.

671 Observations made in the North Atlantic from *RV Michael Sars* in 1910 (Helland-
672 Hansen, 1912) and from *Frithjof* 1910 (Nansen, 1913) showed that the North Atlantic drift brings
673 three branches of warm water towards the eastern part of the Greenland-Scotland Ridge, and two
674 branches cross the ridge. One branch joins the slope current west of Scotland and enters the Arctic
675 Mediterranean through the Faroe-Shetland Channel, where the most persistent, warmest and most
676 saline inflow takes place along the continental slope west of Shetland. This inflow then partly
677 becomes a continuation of the European Slope Current, which is observed at the continental slope
678 west of Spain and may even be present also farther south (Nansen, 1913). The second branch crosses

679 the ridge west of the Faroes and flows first east and then moves south into the Faroe-Shetland
680 Channel, where it splits, one part turning northward into the Arctic Mediterranean and the other
681 circulating around the Faroe Islands. The third branch turns west along the ridge and flows south of
682 Iceland into the Irminger Sea, where the main part continues south along the Greenland slope as the
683 Irminger Current and finally enters the Labrador Sea. A small fraction separates from the stream and
684 flows northward into the Iceland Sea west of Iceland as the North Icelandic Irminger Current,
685 described by Martin Knudsen (1899) based on observations from the Danish *Ingolf* expeditions 1895
686 and 1896.

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

694 At the turn of the century Vilhelm Bjerknes formulated the circulation theorem for a
695 stratified fluid on a rotating earth, which became the basis for the dynamic method by which the
696 geostrophic velocities, arising from the balance between the Coriolis force and the pressure gradient
697 can be determined (Bjerknes, 1898a, 1898b, 1901). How to use the circulation theorem to compute
698 ocean currents was described by Sandström and Helland-Hansen (1903) and Helland Hansen (1905).

699 Helland-Hansen also introduced the TS diagram as a method for describing water masses
700 and stratification in the ocean (Helland-Hansen, 1916). Another advance, based on Nansen's
701 observations on *Fram* that the ice drifted to the right of the wind, was the formulation of the
702 theory of wind driven ocean currents by Vagn Walfrid Ekman (1905).

703

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.

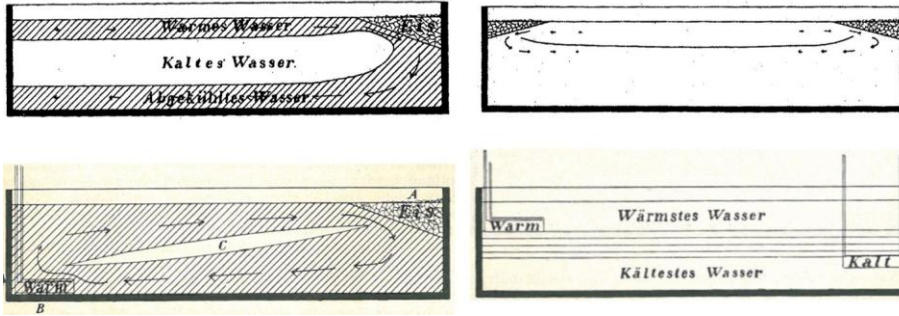
The observations in the Barents Sea close to Novaya Zemlya, by contrast, showed highly saline and cold bottom water, which Nansen interpreted as a result of brine rejection during ice formation. Nansen thus identified two modes of convection; homogenization of the water column by cooling in the open ocean, and the creation of dense saline water by brine rejection on the shelves. Nansen (1912) examined the possible sources of the cold deep water in the North Atlantic and concluded that dense water formed in the Nordic Seas might contribute to the North Atlantic deep water but he expected the main source to be located south of Greenland in the Irminger Sea.

This statement by Nansen may have led to the view that the main transformation of the Atlantic water in the Arctic Mediterranean was into less saline surface water, resulting in too weak exchanges between the North Atlantic and the Arctic Mediterranean (see, e.g., Sverdrup et al., 1942).

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

730 deeper than the cold source. By contrast, if the heat source was above the cold source the water
731 column became stagnant. Since in the ocean the heat and cold sources are both located at the
732 surface, this suggests that a thermally driven circulation is likely to be weak.
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738 Figure 4 The Sandström experiment to simulate the circulation between lower and higher latitudes
739 driven by cooling at high latitudes. The upper panel shows the effect by just cooling, at the northern
740 hemisphere (left) and at both hemispheres (right). The circulation is sluggish. The lower panel shows
741 the circulation with the cold source located above the heat source (left), leading to a vigorous
742 circulation and the situation when the cold source is located below the heat source resulting in
743 stagnant conditions (from Sandström, 1908).

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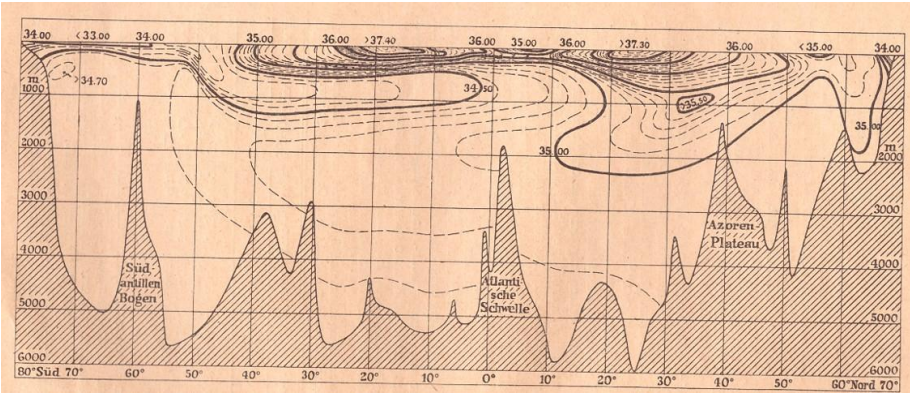
745

746 The density of sea water is determined not only by temperature but also by salinity
747 (and by pressure). This allows for a different circulation mode. In 1755 J.S. von Waitz, a German
748 mining engineer, estimated that evaporation in the Mediterranean Sea leads to a lower sea level,
749 which forces a surface inflow of Atlantic water through Gibraltar Strait. He noted that evaporation
750 removes freshwater but leaves the salt in the water column, which creates a more saline and denser
751 water column in the Mediterranean Sea. This reverses the pressure gradient in depth and gives rise
752 to a dense deep outflow to the Atlantic. Von Waitz was aware of the two-layer exchange through the
753 Bosphorus studied by Luigi Marsigli (1681), which showed less saline water from the Black Sea flowing
754 over a denser deep inflow from the Mediterranean Sea. Von Waitz extrapolated this and his own
755 findings to the world ocean. Heating and evaporation at lower latitudes would increase the salinity
756 and density of the surface water and force it to sink and be replaced at the surface of a flow of
757 colder, less saline and less dense water from higher latitudes. This would create an overturning
758 circulation in the opposite sense to that proposed by von Lenz and Carpenter.

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

771 layers that extend over the entire Atlantic connecting north and south and ultimately the North
772 Atlantic to the rest of the world ocean.
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777 Figure 5. A longitudinal salinity section in the Atlantic showing the presence of the upper low salinity
778 input from the Southern Ocean, the saline North Atlantic deep water and the slightly less saline
779 bottom water from the Antarctic (from Merz and Wüst, 1922).

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781

782 Section 4 The inflows and their forcing

783 The Arctic Mediterranean Sea is dominated by the interplay between the advection of heat, mass
784 and freshwater from lower latitudes and the local, strongly seasonal, forcing. Most heat is carried by
785 the atmosphere, which also provides the largest freshwater input to primarily the Arctic Ocean,
786 mostly as strongly seasonal runoff from the large catchment areas but also as direct local
787 precipitation (Aagaard and Carmack; 1989; Carmack et al., 2015).

788 An oceanic inflow to the Arctic Ocean takes place through [the](#) shallow and narrow
789 Bering Strait and in *the Oceans* H.U. Sverdrup estimated this inflow to 0.3 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3\text{s}^{-1}$)
790 (Sverdrup et al., 1942). Russian work, however, indicated stronger transports and a significant
791 seasonal variability, 1.2 Sv in summer and 0.4 Sv in winter (Maksimov, 1945, Fedorova and Yankina
792 1964). The first direct current measurements made in and north of the strait suggested larger
793 transports, 1.0 to 1.5 Sv, but no seasonal variability was detected (Coachman et al., 1975). The
794 observational work in the strait has continued and later measurements vindicated the early Russian
795 results. The transport varies seasonally with a larger northward transport in summer of 1.2 Sv and a
796 smaller transport, 0.4 Sv, in winter (Coachman and Aagaard, 1981, 1988), and interannually (Itoh et
797 al., 2025).

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

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

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

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

828 The branch of Norwegian Atlantic Current that remains at the slope continues
829 northward to Fram Strait, where it partly merges with the outer Mohn-Knipovich branch. Both
830 branches then split. One stream enters the Arctic Ocean along the continental slope close to
831 Svalbard, and a second stream flows around the Yermak Plateau to rejoin the inner branch farther
832 east. A smaller eastward flow crosses the Nansen Ridge that connecting the Yermak Plateau to the
833 Svalbard slope into the Sofia Deep (Gascard et al., 1995). A substantial fraction of the Atlantic water,

834 however, recirculates westward and does not enter the Arctic Ocean. Instead, it flows south along
835 the East Greenland shelf and slope, as noticed early by Carl Ryder (Ryder 1892). The cooling of the
836 Atlantic water in the Norwegian Sea increases its density sufficiently for it to become overflow water
837 that could contribute to the deep return flow of the North Atlantic overturning circulation.

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

847 The more fundamental question is if the inflow across the Greenland-Scotland Ridge is
848 driven by winds and large weather systems like the North Atlantic oscillation, or if the cooling of the
849 Atlantic water and the convection taking place in the Nordic Seas draw the warm Atlantic water
850 across the ridge in the manner envisaged by Carpenter? This ~~would~~will have a large impact on how
851 the Arctic Mediterranean responds to changes in climate and we return to this question later.

852

853 Section 5 The ice cover and ocean ice atmosphere interactions

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

866 The solid ice cover allows snow to accumulate during winter, which isolates and
867 further reduces the heat loss to the atmosphere and to space. A white snow cover implies a high
868 albedo and when the sun rises in spring most of the shortwave radiation is reflected back to space.
869 This is perhaps the most often mentioned Arctic feedback mechanism. Warmer air, less snow, more
870 absorbed shortwave radiation, warmer air, less snow, etc. However, another mechanism that may be
871 even more important is increased advection of moist air from lower latitudes, which increases the
872 downward longwave radiation and may start melting snow and ice before sunrise (Mortin et al.,
873 2018).

874 During summer the incoming solar radiation increases the air and water temperatures
875 and snow and sea ice start to melt, creating an upper melt water layer that, however, remains close
876 to melting temperature as long as ice is present. If the ice cover is not compact, shortwave radiation
877 penetrates into and through the melt water layer and creates a warmer layer below. This heat is
878 isolated from the ice until autumn, when ice again starts to form and the salinity of the melt water

879 layer increases. These two layers might then merge, which would temporarily slow down the
880 seasonal ice formation. If not, the warmer layer becomes a permanent temperature maximum in the
881 water column, the Near Surface Temperature Maximum (Jackson et al., 2010, 2011).

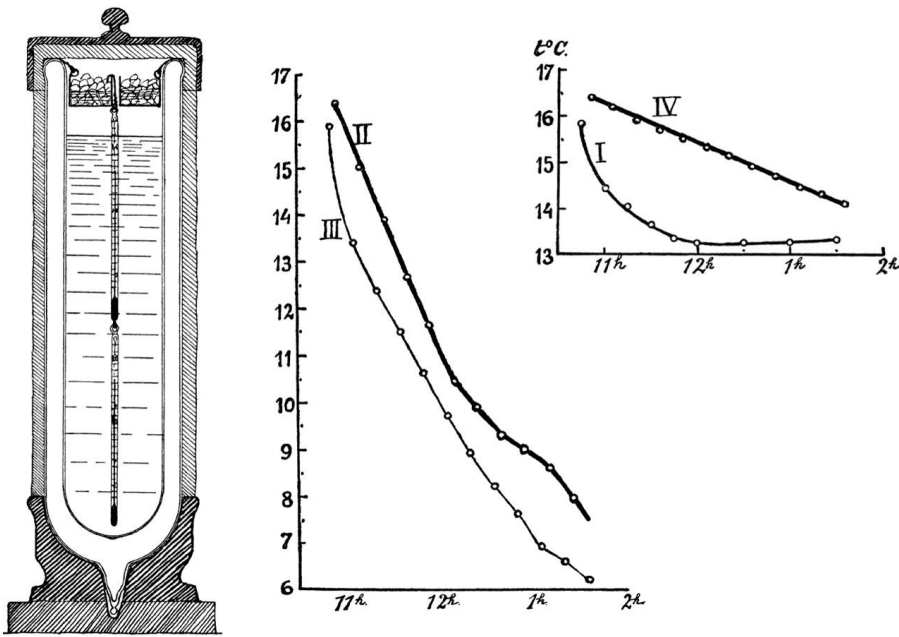
882 A stable seasonal cycle of ice formation and ice melt would eventually lead to an
883 equilibrium thickness of the ice cover. The equilibrium is reached because the ice formation in winter
884 depends more upon the ice thickness than the ice melt in summer. Thinner ice leads to rapid growth
885 in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

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

898 Until now we have only looked at the case when the upper layer temperature is close
899 to the freezing/melting temperature and not considered what happens, when sea ice encounters and
900 melts on warmer water. Nansen (1912) studied this situation by laboratory experiments. A vessel
901 with warm, saline water was cooled from above in three different manners. First a jar with sea ice
902 was held above the water surface and the water was cooled by radiation. Then the jar was allowed to
903 float on the water, cooling the water by conduction. Finally, the ice was added at the surface and
904 coming in direct contact with the ice (Fig. 6). The temperature evolution at the bottom and at the

905 center of the vessel was measured in all experiments. The difference between the three experiments
906 is striking. In the case of just radiative cooling the temperature decrease of the water is small and
907 slow, while the conductive cooling of the water leads to a rapid and large temperature change in the
908 vessel, indicating that thermal convection is established. The experiment where the ice is melting
909 directly at the sea surface is the most revealing. A short rapid cooling takes place but after that the
910 temperature remains essentially constant. The warm water melts the ice, which leads to the
911 formation of a cold, less saline and less dense surface layer that isolates the ice from the warm water
912 below. Sea ice evidently protects itself from heat input from below by creating a stratification that
913 limits the heat transfer.
914

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918

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

925
926

927
928
929 Nansen's experiments were idealized. No turbulence was present in the vessel and the
930 experiments were performed at room temperature. In the ocean wind will generate turbulence,
931 either directly or by driving the ice, which entrains water across the interface into the upper layer.
932 Furthermore, in summer the main sea ice melting occurs at the surface and the melt water layer
933 prevents heat from below to reach the ice. In winter, however, the air temperature is low and the
934 summer melt water has again become ice. As the warm water gets in contact with and melts the ice,
935 an upper, cooler and less saline and less dense layer develops and deepens as warm water is
936 entrained from below.

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

941
942
$$w_e = \frac{2n_o u_*^3}{g[\beta(S_2 - S_1) - \alpha(T_2 - T_1)]} - \frac{\varepsilon B}{g[\beta(S_2 - S_1) - \alpha(T_2 - T_1)]} \quad (1)$$

943

944 Here n_o is a constant usually put to 1.25, u_* is the friction velocity, T_1 , T_2 , S_1 and S_2 are the
945 temperatures and salinities of the upper and lower layer respectively, α is the coefficient of heat
946 expansion and β the coefficient of salt contraction. The sign of α is explicit in the expression. g is the
947 acceleration of gravity, B is the buoyancy input and ε is a parameter put to 1 if the buoyancy input is
948 positive and 0.05 if the buoyancy is negative to take the weak mixing rate of convection into account
949 (Stigebrandt, 1981). If the density is taken only to depend upon salinity, the second term becomes
950 equal to w_e and the input of fresh (melt) water at the surface thus reduces the entrainment by a
951 factor of 2 (Rudels, 1989).

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

$$958 \quad \phi_o = \frac{2\alpha L}{c(\beta S_2 - \alpha \Delta T)} \approx \frac{2\alpha L}{c\beta S_2} \quad (2)$$

959
 960 L is the latent heat of melting and c the heat capacity of sea water and ΔT the temperature
 961 difference between the Atlantic and the melt water layer. (Rudels et al., 1999a). α decreases with
 962 decreasing temperature and Figure 7 shows the total heat loss Q' , the heat loss to ice melt $\phi_o Q'$, and
 963 the heat loss to the atmosphere $(1-\phi_o)Q'$ as functions of ϕ for different ΔT and the corresponding
 964 values of α .

965

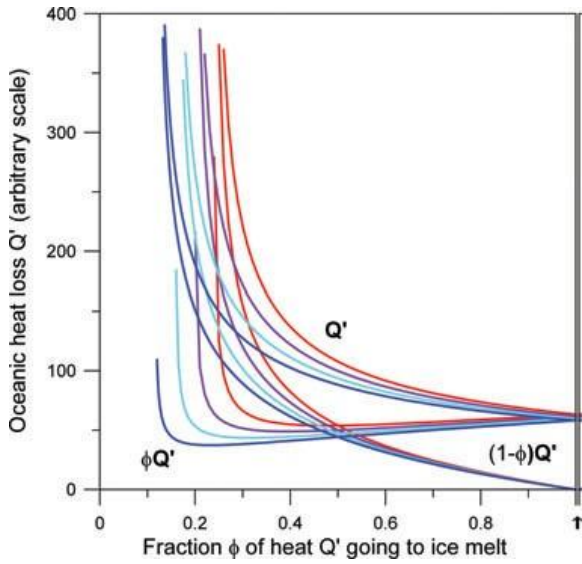


Figure 7. Curves showing the total heat loss Q' , the heat going to ice melt $\phi Q'$ and the heat going to the atmosphere $(1-\phi)Q'$ as function of ϕ , the fraction going to ice melt. Red curves have $\Delta\alpha T = 8^\circ\text{C}$ and $\alpha = 0.781 \times 10^{-4}$, purple curves $\Delta\alpha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

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

$$981 \quad S_1 \approx \frac{S_2}{\left(1 + \frac{2\alpha}{\beta S_2}(T_2 - T_f)\right)} \quad (3)$$

983
 984 T_2 is the temperature of the lower layer and T_f the freezing temperature of the upper layer. The
 985 stable salinity step $\beta(S_2 - S_1)$ is about twice the unstable temperature step $\alpha(T_2 - T_1)$ giving a stability
 986 ratio of 2 at the interface. The problem is the choice of α , which depends strongly on temperature
 987 and as the upper layer deepens and cools α decreases to a minimum at the freezing point. A smaller
 988 α leads to a deeper upper layer with higher salinity.

989 When the freezing temperature is reached, the upper layer has become so deep that
 990 the amount of entrained warm water is not enough to supply heat both to ice melting and to the
 991 atmosphere, and ice again starts to form. The entrained oceanic heat loss, however, is still expected
 992 to be distributed between ice melting and the atmosphere that the ice melting is a minimum. The
 993 stability at the interface now decreases due to ~~the~~ ice formation and brine rejection and the upper
 994 layer eventually becomes dense enough to convect into the deep underlying water (Rudels et al.,
 995 1999a; Rudels, 2021). This situation will be discussed in more detail in section 10 below. Here we
 996 might just add that the fraction ϕ_o of minimum heat going to ice melt is twice the fraction leading to
 997 marginal stability. A fraction smaller than $\phi_o/2$ will not stabilize the water column and it overturns.
 998 This is also evident from the stability ratio $R_p = \beta\Delta S/\alpha\Delta T = 2$ reached when the upper layer has
 999 attained freezing temperature.

1000

1001 Section 6 The Atlantic inflows to the Arctic Ocean

1002 The Atlantic water that enters the Arctic Ocean through Fram Strait soon encounters and starts to
1003 interact with sea ice. In winter the melting sea ice and entrained Atlantic water create a less saline
1004 melt water layer that is eventually cooled to freezing temperature. As a result, an area of open water
1005 often forms north of Svalbard, the “Whalers’ Bay”. Whether heat entrained by mechanical mixing
1006 from the Atlantic layer below is sufficient to melt the ice and create a polynya, or if convection of the
1007 entire upper layer, driven by ice formation and brine rejection after it has reached freezing
1008 temperature, is also required, remains an open question. The Whalers’ Bay normally becomes ice
1009 covered later in winter, either by locally formed ice or by ice advected from the interior of the
1010 Nansen Basin.

1011 The formation of the less saline upper layer implies that part of the Atlantic water that
1012 was cooled in the Norwegian Sea and transformed into high density water, which potentially
1013 ~~contributes~~could contribute to the overturning circulation, instead will leave the Arctic
1014 Mediterranean as a less dense upper outflow. The now stratified Atlantic water column continues as
1015 a boundary current eastward along the Eurasian continental slope and in the following summer sea
1016 ice melting by shortwave radiation creates a seasonal surface melt water layer.

1017 The fate of the water that enters the Barents Sea is more varied and more difficult to
1018 follow. The Norwegian Coastal Current and the main Atlantic inflow branch flow eastward towards
1019 Novaya Zemlya and the heat advected by the Atlantic water is sufficient to keep the southern half of
1020 the Barents Sea free of ice the year around. The heat loss here is large and the Atlantic water
1021 becomes considerably colder than the Atlantic water entering the Arctic Ocean through Fram Strait.
1022 In the southern Barents Sea ice is only formed close to the coast and over the shallow areas west of
1023 Novaya Zemlya, which are mainly covered by water from the less saline coastal current and where
1024 the entire water column is cooled to freezing in winter.

1025 The northern Barents Sea is different. It is shallow, mainly comprising the about 120 m
1026 deep Grand Bank. It connects to the Arctic Ocean between Svalbard and Franz Josef Land with the

1027 deepest trough being the 500m deep Victoria Channel west of Franz Josef Land. The northern
1028 Barents Sea is strongly stratified and seasonally ice covered. The source(s) of the low salinity upper
1029 layer is(are) not obvious. Its characteristics suggest that it is not directly supplied from the Arctic
1030 Ocean, but it might enter from the Kara Sea in the Persey Current, which flows westward south of
1031 Franz Josef Land. Another possibility is that the upper layer is formed in the Barents Sea by sea ice
1032 melting on the warmer Atlantic and coastal waters that move north towards the passage to the Kara
1033 Sea between Franz Josef Land and Novaya Zemlya. The low salinity, cold upper water is eventually
1034 exported from the northern Barents Sea to the Norwegian Sea in the Bear Island Current. The more
1035 saline and denser deep and bottom waters in the northern Barents Sea are supplied by the branch of
1036 Atlantic water that mainly recirculates in the Hopen Deep, but ~~also partly~~ penetrates northward onto
1037 the Grand Bank.

1038 In the 2010s the ice cover in the northern Barents Sea ~~reduced~~was significantly
1039 reduced and the stability between the upper layer and the Atlantic water below weakened. This was
1040 suggested due to an increased inflow of warmer Atlantic water that reduced the stability and melted
1041 the sea ice from below (Lind et al., 2012; Lind et al., 2018). The change would be part of a larger shift
1042 that is presently occurring in the Arctic Ocean, the so called “Atlantification” of the Arctic Ocean,
1043 which implies a reduction of the upper layer thickness and stability, increased shear and a stronger
1044 upward transport of heat from the Atlantic layer to the ice (Årthun et al., 2012; Polyakov et al., 2017;
1045 Årthun et al., 2025). This phenomenon was first reported from the eastern Nansen Basin north of
1046 the Laptev Sea, and the changes observed in the Barents Sea were taken as signs that the same
1047 pattern was expanding. The heat advected from the North Atlantic would in the near future
1048 dominate not only the Norwegian Sea but also the Nansen Basin and the Barents Sea and reduce the
1049 extension of the ice cover (Lind et al., 2018; Polyakov et al., 2020a; 2020b). In the later part of the
1050 2010s the reduction of the ice cover in the northern Barents Sea stopped and it started to recover,
1051 but it is still far from its extent in the 1980s.

1052 Not only the ice cover but also the water mass characteristics in the Barents Sea have
1053 changed considerably during the last decades. In the 1980s the ice cover and the low salinity upper
1054 layer extended farther south and occasionally covered the Central Bank. In some years ice formation
1055 and brine rejection over the bank led to haline convection to the bottom and the formation of a cold,
1056 dense water column over the bank that eventually contributed to dense water formation in and
1057 dense water export from the Barents Sea (Quadfasel et al., 1992; Schauer et al., 2002). In years with
1058 strong inflow of warm Atlantic water the Central Bank, by contrast, became covered by Atlantic
1059 water and no ice and no dense water was formed over the bank. In the 2000s and 2010s the Atlantic
1060 water has been warmer, more saline and denser than in the 1980s. The convection over the Central
1061 bank has been thermal and the created waters have been denser than those formed by brine
1062 rejection in the 1980s.

1063 The dense water formation in the Barents Sea has altered not only over the Central
1064 Bank but also farther east, above the Central Depression and over the shallow areas west of Novaya
1065 Zemlya. The Atlantic water cooled in winter is presently denser than in the 1980s and shows up as
1066 warmer bottom water below a less saline water mass with temperature close to freezing (Lien and
1067 Trofimov, 2013; Dmitrenko et al., 2015). The salinity indicates that the upper layer could be created
1068 by sea ice melting on top of the denser Atlantic water (equation 3 above). This is in contrast to
1069 observations made in the early 1900s, when the water created by ice formation and brine rejection
1070 west of Novaya Zemlya was the densest water formed on the Barents Sea (Knipovich, 1905; Nansen,
1071 1906). Dense water formation by brine rejection still occurs but now it appears to mainly take place
1072 north rather than west of Novaya Zemlya (Lien and Trofimov, 2013).

1073

1074

1075 Section 7 ~~The~~Features of the circulation in the Arctic Mediterranean Sea

1076 Section 7.1 Atlantic water in the Eurasian Basin

1077 The two Atlantic inflow branches ~~meet~~-again meet in the St. Anna Trough in the western Kara Sea.

1078 The Fram Strait branch, flowing as a boundary current along the Eurasian continental slope, follows

1079 the bathymetry southward on the western side of the trough, while the Barents Sea branch enters

1080 the trough from southwest through the passage between Novaya Zemlya and Franz Josef Land. The

1081 properties of the two branches are now different. The Atlantic core of the Fram Strait branch has

1082 been shielded from the surface processes by the less saline upper layer created by ice melting north

1083 of Svalbard and has retained its warm and saline characteristics. The main part of the Barents Sea

1084 inflow branch, by contrast, has been strongly cooled by the atmosphere in the southern Barents Sea

1085 and when it encounters and melts sea ice in the northeastern Barents Sea the salinity and density of

1086 the melt water layer become higher than that created above the Fram Strait branch north of

1087 Svalbard (Equation 3 above).

1088 The part of the Fram Strait branch that flows close to Franz Josef Land enters the

1089 Barents Sea and circulates around Franz Josef Land to return to the Arctic Ocean in the Victoria

1090 Channel, rejoining the boundary current upstream. Most of the Fram Strait branch, however,

1091 interacts in the St. Anna Trough with water of similar density from the Barents Sea, forming a colder

1092 and less saline Atlantic core that remains on the slope and moves eastward between the Kara Sea

1093 shelf and the warmer Fram Strait branch ~~at~~further down the slope. The denser waters formed in the

1094 Barents Sea sink down the eastern flank and along the deepest part of the St. Anna Trough and enter

1095 the boundary current as cold, less saline intrusions beneath the warm core of the Fram Strait branch.

1096 Two of the largest Siberian rivers, Ob and Yenisey, discharge into the Kara Sea, but

1097 most of their runoff does not appear to cross the Kara Sea shelf break. The major part of the river

1098 input instead flows through the Vilkitsky Strait to the Laptev Sea before it, together with runoff from

1099 the Lena, crosses the shelf break and enters the deep Eurasian Basin. This shelf outflow is less saline

1100 and less dense than the surface layer created north of Svalbard and increases the stability of the
1101 upper part of the water column. The Atlantic water becomes more isolated from the ice cover, the
1102 sea surface, and the atmosphere and most of the observed property changes are due to interactions
1103 in the deep between the two inflow branches.

1104 The upper, colder stream follows the shelf break until it reaches Severnaya Zemlya.
1105 Here the slope and shelf narrow and the stream is forced into the basin and starts to interact
1106 isopycnally with the warm core of the boundary current (Rudels et al., 1994; Schauer et al., 1997).
1107 Inversions and thermohaline intrusions are created between the colder, less saline shelf stream and
1108 the warmer, more saline slope current (Fig. 8). Intrusions are observed in all possible background
1109 stratifications; above the temperature maximum where the water column is unstable in
1110 temperature, between the temperature and the salinity maxima where the water column is stable in
1111 both properties, and below the salinity maximum where the water column is unstably stratified in
1112 salinity.

1113

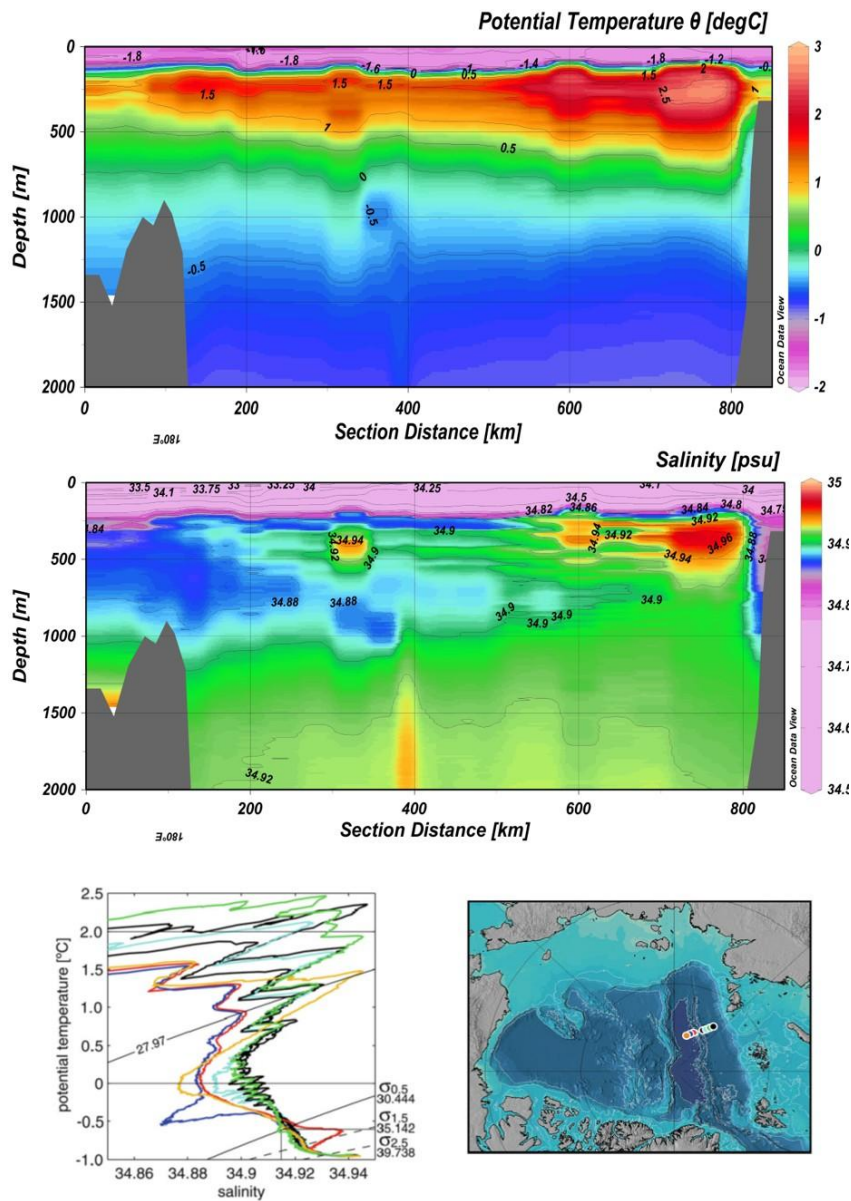


Figure 8. Potential temperature and salinity sections taken in 1996 by *FS Polarstern* from the eastern Kara Sea shelf across the Nansen and Amundsen basins and the Lomonosov Ridge into the Makarov Basin. The $\theta\sigma$ curves are from the same section, showing interleaving as well as eddies in the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

1121 In the Amerasian Basin the Atlantic waters no longer has a salinity maximum and
1122 thermohaline intrusions, the signature feature of the 2000's warming of the Atlantic layer only
1123 persists below the temperature maximum (Carmack et al., 1995; McLaughlin et al., 2009).
1124 Thermohaline intrusions also display a collective, nested structure in T/S space extending fully across
1125 the Arctic Ocean interior (Carmack, 1998; Walsh and Carmack, 20002,2003).

1126 Theoretically, instabilities and interleaving may form spontaneously, if one component
1127 is unstably stratified (Stern 1967; Turner, 1973), but if the background stratification of both
1128 components are stable, finite external disturbances are needed to bring the waters across the front
1129 to generate the property inversions that drive the intrusions. At the slope such disturbances could be
1130 generated by trapped tidal or inertial waves (Toole and Giorgi, 1981). However, the intrusions
1131 penetrate across the front into the opposite water masses only as long as an unstable distribution of
1132 the driving component remains. Once it is exhausted, the interleaving stops (Rudels and Hainbucher,
1133 2020; Rudels, 2021). One question is then: How far from the slope into the basin do the intrusions
1134 influence the lateral mixing between the two streams?

1135 Interleaving structures are observed at the Kara Sea slope but also in the interior of
1136 the Eurasian Basin at least as far as the Gakkel Ridge. However, a core of warm, saline Fram Strait
1137 branch water showing no or weak intrusions is located between these interleaving structures (Fig. 8).
1138 It thus appears that the intrusions formed at the slope do not directly reach to the center of the
1139 warm core. On the basin side of the warm core there is no obvious front that can generate
1140 intrusions. Furthermore, the only source of colder, less saline water in this density interval is the
1141 Barents Sea inflow branch.

1142 One possible explanation for the presence of thermohaline intrusions in the central
1143 Nansen Basin is that the two branches, as they reach the Laptev Sea slope, turn and flow back
1144 towards Fram Strait in the Nansen Basin and over the Gakkel Ridge (Rudels et al., 1994). This implies
1145 that the intrusions do not spread as far as their presence suggests. Instead, the observed interleaving

1146 structures in the interior of the basin are rundown and fossil and are passively advected with the
1147 main circulation. In fact, they could be used as tracers for the circulation (Rudels et al., 1994).

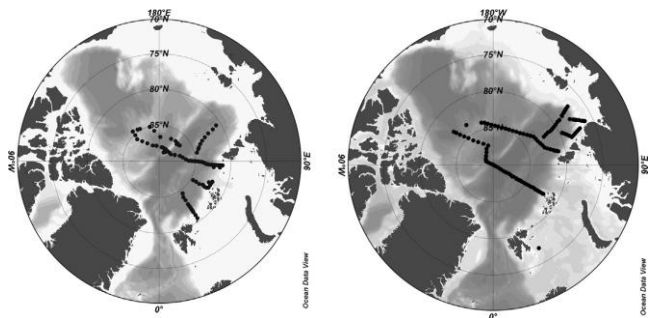
1148 The water column is not only characterized by interleaving and gyre circulation. There
1149 are also several eddies present, which appear to move independently of the gyres, carrying their
1150 original properties into different surroundings (Fig. 8, lower panel).

1151 The interleaving between the two branches leads to cooling of the Fram Strait branch,
1152 but because mass is added the cooling does not necessarily imply a heat loss to the overlying water
1153 and to the ice and atmosphere. The heat has already been lost elsewhere, mainly in the Barents Sea.
1154 Not only is the Atlantic core cooled, the salinity also decreases significantly (Figure 9 below). To
1155 mechanically mix freshwater or low salinity water vertically into the deep is energy consuming, and
1156 the more obvious process is isopycnal mixing between the two branches. The low salinity water
1157 added from the Kara Sea west of Severnaya Zemlya enters, freshens and cools the Atlantic core in
1158 contrast to the low salinity water from the Laptev Sea, which enters as a surface layer and rather
1159 isolates the Atlantic water from the sea surface and the atmosphere.

1160 This is revealed in Figure 9, which shows the heat content (relative to the freezing
1161 point, -1.8°C) and the freshwater content (relative to salinity 34.9) on stations taken on three
1162 sections occupied by *FS Polarstern* in 2007 and 2011 from the Eurasian continental slope into the
1163 Eurasian Basin for six different water masses (Fig. 9). The total height of the columns indicates the
1164 total heat content above 1000 m, but only off the Laptev Sea does the column heights show the total
1165 freshwater content. On the Franz Josef Land and the Severnaya Zemlya sections the freshwater
1166 content in the Atlantic waters is negative and has to be subtracted from the water column heights.
1167 This shows the substantial volume of freshwater that ~~has been~~ added to the deeper Atlantic layers
1168 between the Kara Sea and the Laptev Sea, confirming the importance of the isopycnal mixing
1169 between the branches.

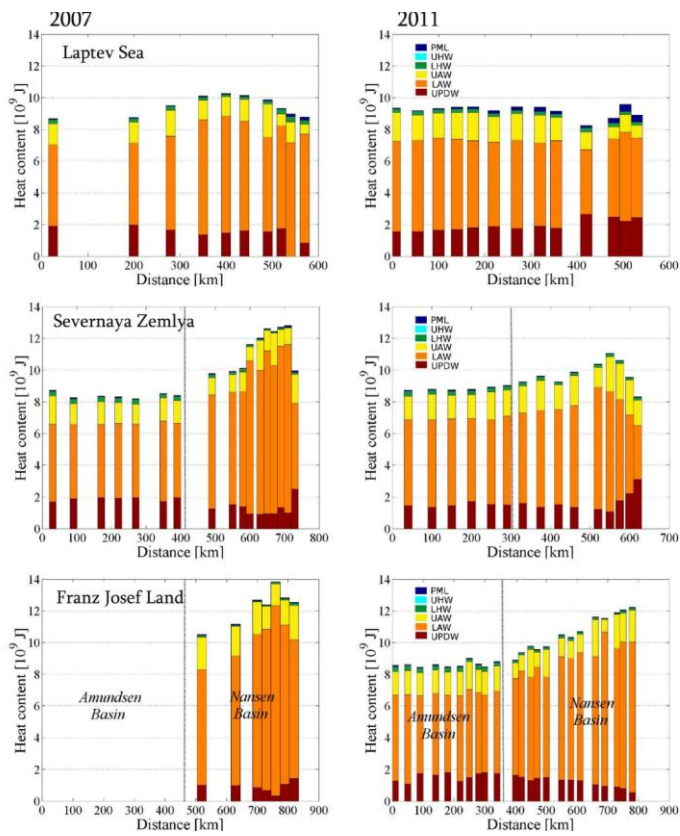
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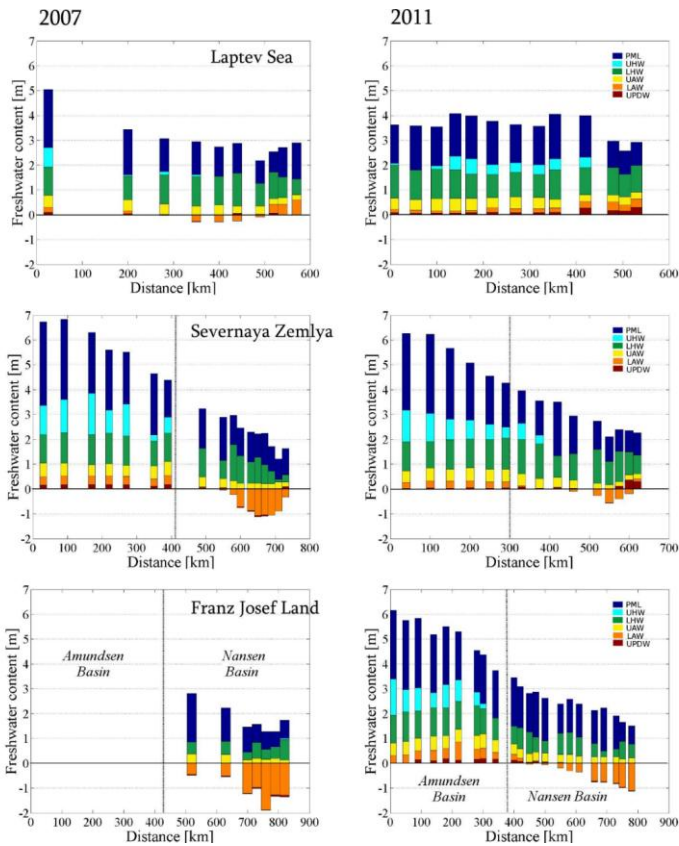


Figure 9. Heat content (central panels) and freshwater content (lower panel) in the upper 1000 m of the water column on stations taken on three sections in the Eurasian Basin in 2007 and almost repeated in 2011. Section positions are shown in the top panel. The heights of the columns indicated the total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

1195 To understand how present, and especially how possible future changes in inflow
1196 strength and characteristics of the Atlantic water might influence the condition in the eastern
1197 Nansen Basin, it is necessary to consider both inflow branches, how they are transformed individually
1198 and how they interact.

1199 Interleaving has vertically alternating salt finger and diffusive interfaces and between
1200 the background temperature and salinity maxima the buoyancy transports through the salt finger
1201 and diffusive interfaces are roughly equal and the property steps diminishes with time but are not
1202 removed. However, if intrusions are created in the upper part of the water column, above the
1203 temperature maximum, the diffusion of buoyancy through the diffusive interfaces is larger than that
1204 through the salt finger interfaces due to the larger property steps created at the diffusive interfaces.
1205 This may lead to overturning of the salt finger interfaces, which transforms the interleaving into a
1206 thermohaline step structure with thick homogenous layers separated by thin diffusive interfaces with
1207 large temperature steps and small stability ratios (Rudels, 2021). Such thick diffusive layers have
1208 been observed in the eastern Nansen Basin above the temperature maximum and they would
1209 contribute to a vertical rather than a lateral heat transport (Polyakov et al., 2012).

1210 The temperature of the Atlantic water in the Nansen Basin was in 1990 observed to
1211 have increased by about 1°C compared to the climatology (Quadfasel et al., 1991). During the 1990s
1212 more changes in the Arctic Ocean water masses were observed. Higher Atlantic water temperatures
1213 were also seen in the Makarov Basin and the Pacific water that previously supplied the upper layer in
1214 the Makarov Basin had retreated from the Lomonosov Ridge to the Mendeleyev Ridge (Carmack et
1215 al., 1993; McLaughlin et al., 1996). This warming trend eventually reached and extended across the
1216 Beaufort Gyre of the Amerasian Basin in early 2000's (McLaughlin et al., 2009).

1217 Also, the runoff normally advected from the Laptev Sea to the Eurasian Basin was
1218 absent, and the more saline melt water layer extended to the sea surface also in the Amundsen
1219 Basin, bringing the Atlantic water closer to the sea surface and the ice cover. By contrast, in the

1220 Canada Basin the upper layer had become thicker and less saline, while the Atlantic water was
1221 pressed down and located at deeper levels (Morison et al., 1998).

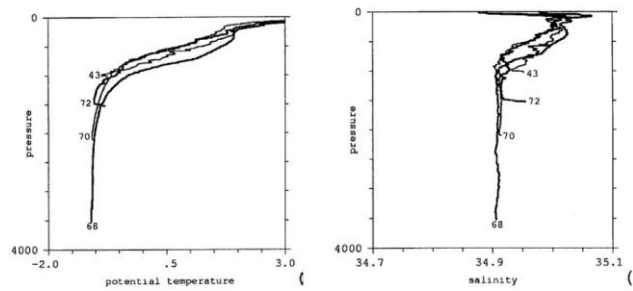
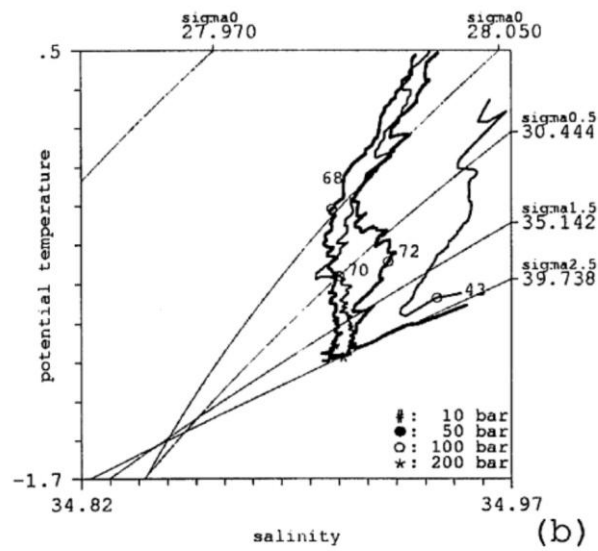
1222 These changes in the Arctic Ocean water column were likely due to changes in the
1223 atmospheric circulation. During the 1990s the North Atlantic Oscillation Index was positive and high,
1224 showing that the Icelandic low pressure system was strong and extended north into the Barents Sea,
1225 perhaps farther. This not only brought more Atlantic water into the Arctic Mediterranean, it also
1226 forced the runoff from the large Siberian rivers to leave the shelves farther to the east, into the
1227 Makarov and Canada basins (Maslowski et al., 2000).

1228 Several later pulses of warm Atlantic water have entered through Fram Strait and
1229 been followed as they moved around in the different gyres in the Arctic Ocean (Polyakov et al., 2005,
1230 2011). Changes have, however, also occurred in the Barents Sea branch. In the 1980s the Barents Sea
1231 branch water that entered the Arctic Ocean was colder, less saline and denser and showed up as a
1232 salinity minimum between 800m and 1200m in the Eurasian Basin (Rudels et al., 2013, see also figure
1233 8). In the 2000s, by contrast, the water that enters and passes over the Barents Sea to the Arctic
1234 Ocean has been warmer, more saline and denser and has instead been identified [in the Eurasian](#)
1235 [Basin](#) as a salinity maximum around 2000 m (Rudels, 2021).

1236 The densest water created by ice formation and brine rejection on the shallow areas
1237 west and north of Novaya Zemlya, however, does not show up clearly in the Arctic Ocean water
1238 column. Its presence has been deduced from an increase in temperature in thin saline bottom layers
1239 at the slope (Rudels et al., 2001). The high salinity is due to brine rejection and accumulation of saline
1240 water at the shelves, and the temperature increase is caused by entrainment of intermediate warm
1241 Atlantic water as the plumes sink into the deep. This process has, however, been followed in more
1242 detail in the western Barents Sea (Fig. 10), where brine enriched dense outflows from Storfjorden in
1243 southern Svalbard sink into the deep Fram Strait, entraining warm Atlantic water (Quadfasel et al.,
1244 1988).

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Figure 10. Potential temperature and salinity profiles and θ_S 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).

1255 There are two different conceptual models for the descending plumes. Shaving-
1256 ~~entraining~~ plumes (Aagaard et al., 1985), where the interfacial stress on the upper part of the plume
1257 causes the plume to gradually lose mass to the surrounding water as it sinks into the deep. The
1258 second model pictures thin plumes passing through and entraining water from the intermediate
1259 layers (Rudels, 1986a). The turbulence is generated by bottom stress and to accommodate the
1260 increased mass transport the plumes do not grow vertically but laterally and their small thickness is
1261 retained.

1262 Shaving plumes are initially fairly thick and carry a substantial amount of dense water
1263 from the shelf into the deep and redistribute it to the ambient water column as they sink. This is a
1264 picture that would fit the denser water descending down the St. Anna Trough, and perhaps also the
1265 dense overflow plumes crossing the Greenland-Scotland Ridge as well as descending Antarctic
1266 ~~slopes~~~~slopes~~ (Carmack et al., 1973-). Entraining plumes, by contrast, would be thin and carry ~~but~~
1267 little but highly saline water from the shelf ~~but~~~~and~~ instead incorporate intermediate water from the
1268 bypassed layers and bring it into the deep. The Atlantic water flowing along the slope thus becomes
1269 transported downwards and transformed into deeper and denser water masses. The increase of
1270 temperature in the bottom layer (Fig. 10) indicates that the boundary plume, initially at the freezing
1271 point, has entrained warm Atlantic water during its descent and the higher salinity relative to the
1272 overlying water column shows that some of the initial salinity excess formed at the shelf still remains.
1273 The profiles and θS curves seen in Fig. 10 thus suggest entraining plumes.

1274

1275 Section 7.2 A possible wind driven circulation

1276 The two Atlantic inflows partly cross the Lomonosov Ridge and enter the Amerasian Basin, partly
1277 recirculate in the Eurasian Basin and return directly to Fram Strait, mostly along prominent
1278 bathymetric features such as the Lomonosov Ridge and the Gakkel Ridge. The forcing of the Atlantic
1279 water circulation in the Arctic Mediterranean deep basins is not well established., but many authors
1280 have presented unified conceptualizations of the circulation of Atlantic ~~Waters~~waters as a boundary-

1281 trapped contiguous flow above the continental slope (cf. Aagaard, 1989; Rudels et al., 1994;
1282 McLaughlin et al., 2002; Aksenov et al., 2011; Bluhm et al., 2020). Here ~~the~~ the approach taken by
1283 Nøst and Isachsen (2003) will be examined.

1284 Nøst and Isachsen (2003) assumed that the circulation is wind driven and computed
1285 the vorticity input at the surface from the NCEP/NCAR reanalysis data (Kalnay et al., 1996, 1997) and
1286 the thermal wind transports from the observed seasonally averaged hydrographic fields. They found
1287 that these did not balance and a Sverdrup circulation, where the added wind torque drives a
1288 meridional transport, is not present in the Arctic Mediterranean Sea. The vorticity added at the sea
1289 surface must then be removed by nonzero bottom velocities and bottom friction.

1290 Nøst and Isachsen suggested either quadratic or linear bottom friction laws to
1291 represent the bottom torque. The applied meteorological forcing corresponds to a vertical velocity of
1292 about 10^{-6}ms^{-1} , and the vertical velocity generated by the bottom torque must be of the same order.
1293 The bathymetry of the Arctic Mediterranean consists of several basins that display closed
1294 topographic contours H and also closed geostrophic contours f/H , H being the depth and f the
1295 Coriolis parameter. Assuming that the bottom circulation follows the geostrophic contours and that
1296 the width of the bottom streams is equal to the width of the continental slopes, the bottom
1297 velocities can be estimated to 10^{-1}ms^{-1} , which is close to the observed bottom velocities (Newton and
1298 Sotirin, 1997; Nøst and Isachsen, 2003). The picture is then of an almost barotropic deep circulation
1299 in the Arctic Mediterranean that largely follows the f/H contours and that the wind forcing is
1300 balanced by the bottom torque.

1301 The neglect of thermodynamics, except that present in the observed and used density
1302 field, prevents the model from addressing the changes in the stratification that take place in the
1303 Arctic Mediterranean. Aaboe and Nøst (2008) tried to remedy this limitation by introducing a
1304 variation in bottom density along the f/H contours. The bottom density is highest in the Nordic Seas
1305 and lower in the Arctic Ocean and especially in the Canada Basin. This generates a cyclonic flow in
1306 the Nordic Seas, which weakens and eventually changes to anticyclonic in the Arctic Ocean. This

1307 change can be explained by a shift from a strong barotropic circulation in the Nordic Sea to a larger
1308 contribution of the baroclinic part of the circulation (Aaboe and Nøst, 2008). No thermodynamics is
1309 included in the model that can explain the bottom density changes, but one possibility could be that
1310 the input of shelf water and the Pacific inflow at the surface ~~push~~pushes the boundary current to
1311 deeper levels and ~~replace~~replaces the water column at the chosen f/H contour with a less dense one
1312 from higher up on the slope.

1313 Nøst et al. (2008) made a theoretical and experimental study of two connected basins
1314 with sloping bottom and with one basin driven by either cyclonic or anticyclonic forcing. They found
1315 that cyclonic forcing at one basin also made the second basin exhibit a cyclonic circulation. By
1316 contrast, a weak anticyclonic forcing in one basin creates an anticyclonic circulation in the second
1317 basin. However, when the anticyclonic forcing increases the circulation in the second basin shifts to
1318 cyclonic. It thus appears that a cyclonic circulation is more stable than an anticyclonic one. Another
1319 interesting detail is that cyclonic forcing in one basin can also drive a cyclonic circulation in an
1320 adjacent basin. This implies that an atmospheric forcing in the Nordic Seas could drive ~~the~~a cyclonic
1321 circulation along the geostrophic contours also in the Arctic Ocean. If the cyclonic wind forcing in the
1322 Subpolar gyre can drive the Atlantic inflow to the Arctic Mediterranean is then another interesting
1323 possibility.

1324

1325 Section 8 The ~~circulation~~conditions in the Amerasian Basin

1326 8.1 The upper layers and the ice cover

1327 East of the Laptev Sea and the New Siberian Islands the nature of the circulation changes. It is no

1328 longer dominated by the two Atlantic inflows but by freshwater input from the low salinity Pacific

1329 water and from rivers and continental runoff. The inflow through Bering Strait has since the

1330 beginning of this century been monitored by direct current measurements and the estimates

1331 ~~indicates~~indicate an increase in ~~both~~ volume and freshwater transports as well as in heat transport

1332 (Woodgate and Aagaard, 2005; Woodgate et al 2012; Woodgate et al., 2015). The ~~increase~~increased

1333 heat and freshwater transports might largely be due to the observations also of the mainly seasonal

1334 Alaskan Coastal Current that carries freshwater from the Yukon river discharging into the Bering Sea

1335 (Woodgate and Aagaard, 2005). The mean yearly volume transports are presently estimated to 1.2 Sv

1336 and about one third the freshwater input (relative to 34.80) to the Arctic Ocean takes place through

1337 Bering Strait (Dickson et al. 2007).

1338 The resulting freshwater storage confines the atmospheric forcing and the oceanic

1339 response to the upper part of the water column. The mainly wind driven circulation is centered

1340 below the atmospheric high pressure cell above the Beaufort Sea in the Canada Basin. The

1341 anticyclonic wind field forces the less dense surface waters towards the center of the Beaufort Sea,

1342 leading to a higher sea level and to an anticyclonic geostrophic circulation, the Beaufort Gyre, in the

1343 upper few hundred meters.

1344 In the Eurasian Basin, by contrast, the low pressure cell extending from Iceland into

1345 the Barents Sea creates a cyclonic circulation that drives Atlantic water from the Norwegian Sea into

1346 the Arctic Ocean and forces much of the continental runoff from Eurasia into the Amerasian Basin.

1347 The two gyre systems are brought together and interact in the Transpolar Drift, where

1348 the Siberian branch originating in the Laptev Sea flows towards Greenland and Fram Strait in the

1349 Amundsen Basin parallel to the northward flowing part of the Beaufort Gyre. Waters are exchanged

1350 between the two gyres and the denser Siberian branch intrudes into and supplies water to the

1351 deeper levels of the Beaufort Gyre (Morison et al., 2012). Both streams transport sea ice and
1352 especially the Laptev Sea has been considered an important formation site for the sea ice exported
1353 from the Arctic Ocean.

1354 The existence of the Transpolar drift was demonstrated by the drift of *Fram* 1893-
1355 1896. Roald Amundsen tried to repeat the drift with *Maud*, 1918-1925, but by starting from the East
1356 Siberian Sea farther to the east, he hoped to have a greater possibility to drift over the North Pole.
1357 This plan did not work. *Maud* remained in the East Siberian Sea trapped by the ice and did not pass
1358 beyond the shelf break. The main source of the Transpolar drift thus appears to be located west of
1359 the New Siberian Islands, while the East Siberian Sea in contrast to the Laptev Sea receives a
1360 considerable amount of sea ice from the central basins and more acts as an ice sink than an ice
1361 source for the inner Arctic Ocean. This is partly confirmed by Zacharov (1976), who estimated the ice
1362 export from Siberian shelves and concluded that the Laptev Sea ice export was about ten times as
1363 large as that from the East Siberian Sea.

1364 The ice cover in the East Siberian Sea has diminished drastically in recent years and has
1365 now changed to seasonal from being present almost the year around. This is but one indication of
1366 the changes taking place in the Arctic Ocean ice cover. Such changes ~~are~~may even ~~affecting~~affect
1367 structures deep within the water column (Bretones et al., 2022). The decline in sea ice extent has
1368 since the beginning of the satellite observations in the 1970s been followed almost in real time, but
1369 information about sea ice thickness is more difficult to obtain. Not until upward-looking sonar
1370 observations were made from the submarine USS Pargo in 1993 could sea ice thickness
1371 measurements be compared with those made about 30 years earlier. The thickness of the Arctic
1372 Ocean ice cover had decreased by 40%, from an average thickness of about 3 m to 1.8 m (Rothrock et
1373 al., 1999).

1374 The thickness reduction continued and accelerated in the following decades (Kwok et
1375 al., 2009) and a major shift in the sea ice distribution was observed. The earlier dominance of thick
1376 and deformed multiyear ice changed ~~in to~~into a stronger presence of thinner first year ice (Kwok et

1377 al., 2009; Sumata et al., 2023). This shift appears to have taken place in the first decade of this
1378 century (Kwok et al., 2009) and Sumata et al. (2023) relate the change to the strong sea ice export
1379 events that occurred in 2005 and 2007. Once the main storage of multiyear ice had been removed
1380 the drift speed of the sea ice increased and its residence time decreased, which prevents the buildup
1381 of a large volume of multiyear ice in the Arctic Ocean.

1382 The effects of these changes were clearly demonstrated by the ship *Tara*, which in
1383 2008 within the EU-DAMOCLES (Developing Arctic Modeling and Observing Capabilities for Long-
1384 term Environmental Studies) program repeated the drifts of *Fram* and *G.Sedov* in only 18 months
1385 compared to 3 years for the earlier drifts, showing that thinner ice moves more quickly than the thick
1386 ice cover in the past. The reduction in ice cover thickness and extent ~~the~~ does not appear to have
1387 significantly changed the sea ice export to the Nordic Seas through Fram Strait (Sprensen et al., 2009).

1388 The ice grows as it drifts and thinner ice grows more quickly, and the reduced ice
1389 thickness is also compensated by the more rapid flow. However, the storage of freshwater as ice in
1390 the Arctic Ocean has declined significantly, and the seasonal ice formation manages to supply the ice
1391 export but cannot rebuild any large reservoir of especially older, thicker ice within the Arctic Ocean.
1392 Some of the more dramatic reductions in the ice cover have been connected with atmospheric
1393 circulations and winds favoring a large ice export through Fram Strait as in 2007, the year of the
1394 second lowest summer ice extent on record.

1395 The loss of sea ice has not resulted in a reduction of the total freshwater storage in the
1396 Arctic Ocean. The total freshwater volume has increased by 4500km³ and the liquid freshwater that
1397 has accumulated in the Beaufort Gyre has increased the freshwater storage there by 5000km³,
1398 between 2000s and the 1980s and 1990s. This more than compensates for the about 3500km³ loss of
1399 sea ice (Haine et al., 2015).

1400

1401 8.2 The Beaufort Gyre

1402 The circulation in the Beaufort Gyre is complex (McLaughlin et al., 2009; Timmermans and Toole

1403 2023; ~~Timmermans~~Timmermans et al., 2025). The anticyclonic winds collect the low salinity upper
1404 water in the center of the gyre. As a result, the Beaufort Gyre is the major freshwater storage
1405 reservoir within the Arctic Ocean (Aagaard and Carmack, 1989). The volume of freshwater within this
1406 reservoir is subject to interannual variability, and poses a risk of catastrophic discharge into the
1407 convective gyre of the northern North Atlantic (Proshutinsky et al., 2009; ~~2015~~2015; 2019). This
1408 accumulation is balanced by instabilities in the deeper layers that shed eddies into the surroundings
1409 (Manucharyan and Spall, 2016). However, the balance between these two processes is yet not
1410 reached, and the present storage of about 20 m of freshwater relative to a salinity of 34.8 could
1411 increase to more than 30 m (Manucharyan and Spall, 2016; Manucharyan et al., 2016). Another,
1412 additional, process has therefore been suggested, the ice-stress governor, resembling the valve that
1413 in the 19th century prevented steam engines from running amok (Meneghello et al., 2018).

1414 The accumulation of low salinity water by the wind in the center of the gyre creates a
1415 baroclinic density field that drives an anticyclonic geostrophic circulation around the gyre. If the wind
1416 stress on the ice surface is taken up by the internal stresses in the ice cover, the ice cover is no longer
1417 forced by the wind. Instead, its motion ceases and it begins to retard the geostrophic circulation of
1418 the water below, which in turn reduces the slope of the isopycnals and the accumulation of
1419 freshwater. This process should then act together with the instability mechanism to reduce the
1420 freshwater storage in the Beaufort Gyre (Meneghello et al., 2018). However, if the ice cover becomes
1421 less compact, or perhaps disappears, the ice-stress governor ceases to function and more freshwater
1422 would again be collected in the Beaufort Gyre (Doddridge et al., 2019).

1423 With more freshwater accumulated in the Beaufort Gyre the stability increases and
1424 the upper layer becomes isolated from the bathymetry. The balance between the input of vorticity at
1425 the surface and its removal by bottom stress, as proposed by Nøst and Isachsen (2003), might then
1426 disappear and another balance is needed. The possibility of a Sverdrup balance has usually been
1427 discarded because of the small β at high latitudes. Yang et al. (2015) examined the situation for a
1428 circular basin in high latitudes. They found that in the case of no boundaries the effects of the

1429 difference in β between latitudes was less pronounced compared to the case when a western
1430 boundary was artificially introduced. They concluded that the vorticity input from the anticyclonic
1431 wind stress curl would in the gyre be balanced by the time mean divergence of the eddy vorticity flux
1432 and along the western boundary by a frictional torque.

1433 No solid western boundary is, however, present near the Beaufort Gyre, but there is a
1434 northward flowing boundary current that is fed both from the east and from the west, the Transpolar
1435 drift. The Beaufort Gyre is a unique feature in the world ocean. It is forced by anticyclonic winds,
1436 creating the density stratification and the gyre circulation. There should be an, albeit weak,
1437 southward Sverdrup transport, which could continue south through the Canadian Arctic Archipelago
1438 but also ~~could~~ partly could feed an inertial current flowing northward on the western side of the
1439 gyre. The increase in planetary vorticity would then be balanced by an increase in negative relative
1440 vorticity in the boundary current. The flow would intensify as long as f increases, but when the
1441 highest latitude is reached, f starts to decrease and the boundary current begins to diverge, reducing
1442 its relative vorticity. This is different from mid latitudes, where a purely inertial boundary current
1443 cannot exist because the boundary current must shed its relative vorticity, when it leaves the
1444 boundary, which only can ~~take place~~ be accommodated by friction (Vallis, 2017).

1446 8.3 Vertical exchanges, diffusive interfaces and the circulation in the Atlantic layer

1447 In the Amerasian Basin the input of low salinity water from the shelves, and the Pacific inflow that
1448 create a temperature maximum around 75m in summer and form a temperature minimum between
1449 200m and 300m in winter, further isolate the Atlantic water from the energy input at the sea surface.
1450 Variability in the inflow of summer water, which lies immediately below the mixed layer, has been
1451 observed to influence the ice cover (Shimada et al., 2006). The underlying temperature minimum
1452 also involves an increase of salinity, suggesting that in winter ice has formed in the Bering and
1453 Chukchi seas, and that the bottom water salinity has increased by brine rejection. This is in
1454 agreement with the mechanism for creating the cold Arctic Ocean halocline proposed by Aagaard et

1455 al. (1981), which isolates the deeper layers from the surface processes. The cold, saline water also
1456 has high nutrient content, indicating nutrient regeneration ~~of nutrients~~ from the bottom sediments
1457 (Jones and Anderson, 1986).

1458 The thermocline between the Pacific winter water and the temperature maximum of
1459 the Atlantic layer displays step structures and homogenous layers, suggesting that heat is
1460 transported upwards by double diffusion. The individual diffusive layers ~~and steps, about 5 m thick,~~
1461 appear to be coherent over ~~several~~ 800 kilometers, giving a horizontal to vertical ratio of about
1462 ~~10000000000~~ 200000 to 1 (Timmermans et al., 2008). However, over the same distance large temperature
1463 differences are present in both the upper and the lower part of the thermocline. The vertical
1464 diffusive heat flux reduces the temperature of the Atlantic core, but the lost heat becomes trapped
1465 in and warms the temperature minimum of the Pacific winter water. The temperature in the central
1466 part of the thermocline, however, remains more or less constant, suggesting that here the heat
1467 fluxes through the lower and upper diffusive interfaces are balanced. This implies that the
1468 thermocline and the diffusive layers are advected over the basin on time scales of several years. The
1469 layers are coherent in time rather than in space.

1470 The stability ratios $\beta S/\alpha T$ at the interfaces are large, 3-4. This implies that parcels
1471 convecting from the interfaces can homogenize the existing layers but could not have created the
1472 initial layers. The layers, or embryos of the steps and layers might then be present already as the
1473 Atlantic water enters the Amerasian Basin and the larger number of layers and interfaces compared
1474 to the Eurasian Basin ~~would~~ could be due to a decreased buoyancy anomaly of the rising and sinking
1475 plumes. They are then no longer able to stir and homogenize the entire layer but only the part close
1476 to the interfaces, and new interfaces are created at mid-depth.

1477 The Atlantic water that returns from the Amerasian Basin lies too deep to exit through
1478 the Canadian Arctic Archipelago. Instead, it appears, at least partly, to leave the North American
1479 continental shelf and slope and move northward as a boundary current along the Lomonosov Ridge.
1480 There it passes through gaps in the ridge from the Makarov to the Amundsen Basin and then returns

1481 south towards Greenland (Björk et al., 2010). Some Atlantic water may also cross the Lomonosov
1482 Ridge at the continental slope directly to the Amundsen Basin.

1483

1484 8.4 The deep waters in the Arctic Ocean

1485 In the mid-20th Century the commonly accepted view was that the deep and bottom water source in
1486 the Arctic Mediterranean was located in the Greenland Sea and that cold, dense deep water
1487 penetrated into the Arctic Ocean through Fram Strait and its temperature increased with distance
1488 from the source (Wüst, 1941). However, on the US expedition Ski-jump 1951 and 1952 L.V.
1489 Worthington noticed that the deep water temperatures below about 1600 m in western Arctic
1490 Ocean, away from Fram Strait, remained almost constant around -0.5°C , significantly higher than in
1491 the Greenland Sea. Worthington suggested that the higher temperature was caused by a submarine
1492 ridge that prevented the deepest and coldest water from spreading to the entire Arctic Ocean
1493 (Worthington, 1953). Unknown to Worthington this ridge, the Lomonosov Ridge, had been
1494 discovered by the Soviet airborne high-latitude expeditions in 1948.

1495 It was later noticed that not only the deep temperatures but also the deep salinities
1496 were higher in the Amerasian Basin beyond the Lomonosov Ridge than in the Eurasian Basin
1497 (Aagaard, 1980). This implies that the deep Arctic Ocean is not stagnant and that water mass
1498 renewals must take place. The situation was akin to the one Nansen had encountered with the
1499 erroneous, too high, salinity measurements from *Fram*. Nansen had initially suggested that these
1500 salinities could be due to brine enrichment of the bottom waters on the shallow shelves and a
1501 subsequent sinking of the dense shelf water down the continental slope (Nansen, 1906). An idea that
1502 he later abandoned, when he found that the deep salinities measured north of Svalbard agreed with
1503 those observed in the Greenland Sea (Nansen, 1915). Now the Arctic Ocean was suddenly recognized
1504 as an additional active source for the deep water in the Arctic Mediterranean and Aagaard et al.
1505 (1985) and Rudels (1986a) proposed a connected system, where the Greenland Sea supplies the

1506 colder, less saline end member and the Arctic Ocean the saline, warmer end member and the two
1507 sources communicate through Fram Strait.

1508 The water columns in the Eurasian Basin and in the Amerasian Basin differ. The
1509 Eurasian Basin is warmer and more saline in its upper part while the Amerasian Basin is warmer and
1510 more saline in the deep. The changeover occurs at different depth levels for temperature and
1511 salinity. The Amerasian Basin becomes warmer than the Eurasian Basin below 900 m, well above the
1512 sill depth of the Lomonosov Ridge, but it becomes more saline first at 1400 m, closer to the sill depth,
1513 but still in a depth range, where the different basins communicate with each other.

1514 These differences must be due to slope convection and the entrainment into the
1515 sinking boundary plumes. The colder deep water in the Eurasian Basin could be explained by the
1516 entrainment of much colder ambient water as the plumes sink through the deeper part, while the
1517 more saline characteristics of the Amerasian Basin may be caused by the less dense upper part of the
1518 water column, which requires the initial shelf contributions to be much more saline to reach the
1519 deepest layers than those descending from the Eurasian shelves.

1520 Below about 1500 m to 2000 m the deep waters in the different basins communicate
1521 less freely and they have acquired their different physical, chemical and biological characteristics
1522 (Bluhm et al., 2016). The deep water columns in the Nansen, Amundsen and Canada basins are
1523 similar. The temperature decreases and the salinity increases until about 1000 m from the bottom,
1524 where a temperature minimum is encountered. Below this level the temperature increases slightly
1525 and the salinity continues to rise until a homogenous, 600 m to 800m thick bottom layer is
1526 encountered. The temperature increase and the homogenous bottom water have been explained by
1527 geothermal heating, which could homogenize the bottom layer by convection from below and
1528 perhaps eventually remove the temperature minimum (Timmermans et al., 2003; Björk and Winsor,
1529 2006; Carmack et al., 2012; Arroyo et al., 2025). The ~~presence~~existence of a temperature minimum in
1530 the three basins is not obvious. In the Canada Basin it can be explained by advection of deep water

1531 from the Makarov Basin across the Alpha and Mendeleyev ridges, but no such source can be
1532 identified in the Nansen and Amundsen basins.

1533 The initial temperature of the shelf water is at freezing and practically constant while
1534 its initial salinity depends on the amount of released brine. The advective water that is entrained into
1535 the descending plumes might, however, vary in time. A colder, and denser, advective water column
1536 would make the descending plumes colder and denser and the coldest plumes would penetrate the
1537 deepest. Combined with geothermal heating of the bottom layer formed by the most saline plumes a
1538 deep temperature minimum could then evolve.

1539 The Makarov Basin is different. No deep temperature minimum is present and the
1540 salinity increase with depth stops about 1000 m from the bottom, while the temperature continues
1541 to decrease until 500 m above the bottom. Below this level a homogenous bottom layer is found.

1542 This feature was explained by Jones et al. (1995) ~~by~~^{as} a spillover across the Lomonosov Ridge of
1543 colder, less saline deep water from Amundsen Basin, which could sink to the bottom, aided by the
1544 higher compressibility of colder water. However, when the deepest sill at the Lomonosov Ridge was
1545 reached and studied by *IB Oden* in 2005, no such overflow was found. Instead, warm and saline deep
1546 water from the Makarov Basin was observed entering the Amundsen Basin (Björk et al., 2007). If
1547 there is a spillover from Amundsen Basin to the Makarov Basin, it must be intermittent.

1548

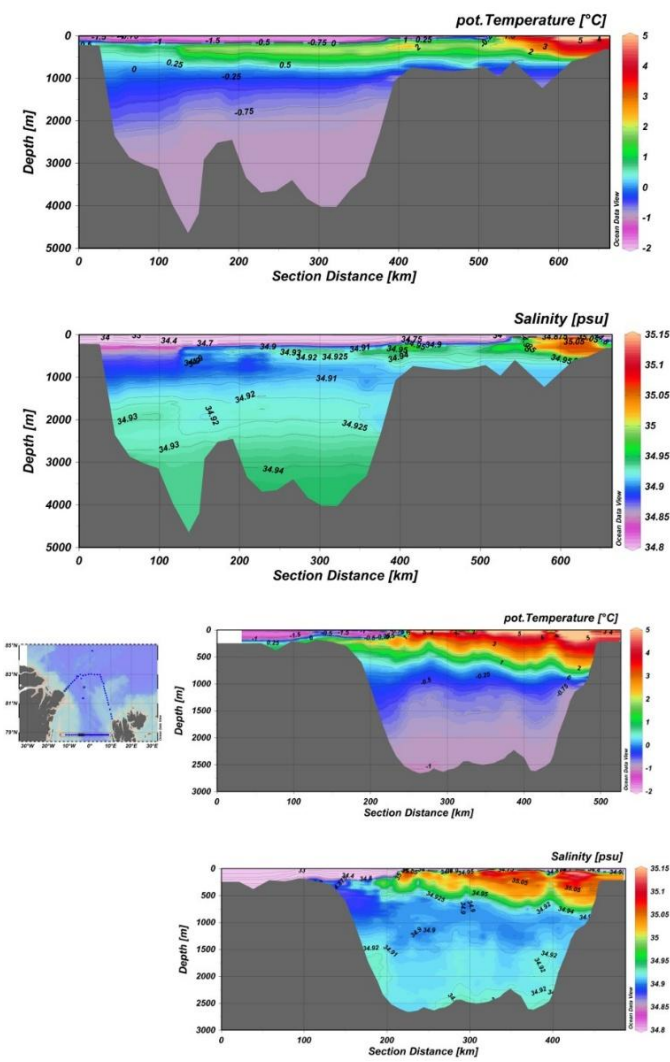
1549 Section 9 Fram Strait.

1550 The downstream conditions of the inflow to the Arctic Ocean through Fram Strait and the upstream
1551 conditions ~~for of~~ the return flows to the strait are shown in the northern section in Fig. 11. The inflow
1552 involves not only Atlantic water, which mainly enters over the shelf north of Svalbard, but also
1553 denser, deeper lying waters, which cannot pass directly eastward and have to circulate along the
1554 Nansen Ridge and the northern slope of the Yermak Plateau before they can join the shelf inflow at
1555 the continental slope east of Svalbard. Farther north waters returning ~~to~~ towards Fram Strait in the
1556 northern Nansen Basin, over the Gakkel Ridge and in the Amundsen Basin dominate. ~~Both Greenland~~
1557 ~~Sea Deep Waters and Eurasian Basin Deep Waters above Fram Strait have warmed in recent years,~~
1558 ~~such that the Greenland Sea is now a source of warming to the interior Arctic Ocean (Karam et al.,~~
1559 ~~2024; Song et al., 2025). Both Greenland Sea Deep Waters and~~ Both the Greenland Sea Deep Waters
1560 and the Eurasian Basin Deep Waters above Fram Strait have warmed in recent years, such that the
1561 Greenland Sea is now a source of warming to the interior Arctic Ocean (Karam et al., 2024; Song et
1562 al., 2025).

1563 The Arctic Atlantic water returning in the Eurasian Basin is warmer and more saline
1564 than the Arctic Atlantic water that has circulated through the Amerasian Basin and approaches Fram
1565 Strait along the Greenland slope. The water that enters the Arctic Ocean over the Barents Sea is
1566 identified as a colder, less saline layer at 1000 m depth, but it is difficult to separate the intermediate
1567 waters recirculating ~~directly~~ in the Eurasian Basin from those that have passed through the
1568 Amerasian Basin, except by location. The deep water from the Amerasian Basin shows up as a salinity
1569 maximum around 2000m at the Greenland slope, and the colder and denser Eurasian Basin deep
1570 water fills the deeper layers on the northern section (Fig. 11).

1571

1572



1573

1574

1575 Figure 11. Two potential temperature and salinity sections, one taken from Svalbard along the
1576 Nansen Ridge and over the Yermak plateau into the Nansen Basin and then westward to Greenland
1577 and the second along the sill in Fram Strait showing the exchanges between the Arctic Ocean and the
1578 Nordic Seas (sections taken by *FS Polarstern* 2004).

1579

1580

1581 On the southern section, taken along 78°50'N in Fram Strait (Fig. 11), there is an inflow
1582 of mainly Atlantic water in the West Spitsbergen Current but also of intermediate and deep waters
1583 from the Nordic Seas. The deep waters in general are slightly less saline in Fram Strait than farther
1584 north in the Nansen Basin. The central part of the strait exhibits a strong recirculation of Atlantic
1585 water towards the west. The recirculation is largely carried by barotropic and baroclinic eddies that
1586 appear to move westward towards Greenland. The main southward flow from the Arctic Ocean takes
1587 place in the west in the East Greenland Current. The recirculation in Fram Strait is strong and only a
1588 narrow space is available for the outflowing water between the recirculating Atlantic water and the
1589 Greenland continental slope.

1590 The East Greenland Current is more baroclinic than the West Spitsbergen Current and
1591 carries several different water masses. The upper layer is dominated by low salinity waters
1592 comprising river runoff, occasional contributions from the Pacific inflows, and halocline water
1593 created from ice melted on the warm Atlantic water entering north of Svalbard. The cold, low salinity
1594 water flows above warm recirculating Atlantic water that penetrates into the East Greenland Current
1595 above the transformed Arctic Atlantic water that is returning along the Greenland continental slope.
1596 The warmest Arctic Atlantic water derives from the Fram Strait branch that mainly remains in the
1597 Eurasian Basin while the colder contributions are supplied by the Barents Sea inflow branch that
1598 circulates around the Makarov and Canada basins. The Barents Sea inflow is transformed into an
1599 intermediate water mass with distinct doubly stable θS characteristics, salinity increasing and
1600 temperature decreasing with depth.

1601 At the western slope two outflows of the Arctic Ocean deep waters can be identified.
1602 A warm, saline core centered around 1800m depth, which ultimately derives from the Amerasian
1603 Basin, and a colder core with similar salinity located at sill depth, indicating an outflow of Eurasian
1604 Basin deep water. The comparatively warm and saline characteristics of the Arctic Ocean deep
1605 waters indicate that they are created by slope convection as shelf water made dense by ice
1606 formation and brine rejection sinks down the slope entraining ambient warmer water. The salinity

1607 then decreases and the temperature increases and the plumes continue to sink until their density
1608 matches that of the surrounding water column. The deep and bottom water of the Arctic Ocean thus
1609 become warmer and more saline than deep water created by open ocean convection. The
1610 entrainment into the boundary plumes transports Atlantic water downward in the water column and
1611 its density increases as it mixes with the cold, saline shelf outflow. The interaction with sea ice
1612 formation and sea ice melting thus separates the inflowing Atlantic water into both less dense and
1613 denser waters.

1614 Fram Strait is the only deep passage to the Arctic Ocean and most of the entering
1615 Atlantic water must exit through Fram Strait. The low salinity waters can, however, leave the Arctic
1616 Ocean either through Fram Strait or through the narrow and shallow passages of the Canadian Arctic
1617 Archipelago. The largest runoff enters the Kara and Laptev sea shelves and since the shelf outflow
1618 from the Laptev Sea to the Amundsen Basin has salinity between 32 and 33 the runoff must mix with
1619 an advective saline water mass, and the only one available on these shelves is the water of the
1620 Norwegian Coastal current that enters the Barents Sea and continues eastward to the Kara Sea and
1621 Laptev Sea shelves.

1622 The Chukchi shelf is flooded from behind by the Pacific inflow, which also supplies
1623 saline water to the East Siberian Sea, where, in addition, upwelling of saline water from the Makarov
1624 Basin takes place. The melting of sea ice on the entering Atlantic water, either north of Svalbard or in
1625 the Barents Sea, creates the densest upper layer water and the densest water that can pass through
1626 Nares Strait, the deepest passage in the Canadian Arctic Archipelago (Rudels et al., 2004; Díckson et
1627 al., 2007).

1628 The major driving forces behind the exchanges through Fram Strait are not identified
1629 with certainty, but the largest transports take place at the slopes, which could indicate either a
1630 barotropic wind driven flow along geostrophic contours as ~~that~~ suggested by Nøst and Isachsen
1631 (2003), but also baroclinic transports of mainly less dense upper layer waters along the coastal
1632 boundaries (Werenskiold, 1935, Stigebrandt, 1981), or a combination of both. The sea level in the

Arctic Ocean is higher than in the Nordic Seas, which favors a southward flow in the East Greenland Current. It could also obstruct the northward flow in the West Spitsbergen Current and force the westward recirculation in the strait.

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

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

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

1659 greatest differences were seen between the Greenland and the Svalbard shelves and slopes, while
1660 the water flowing north and south in the central part of the strait were similar, indicating a strong
1661 recirculation.

1662 The northward and southward volume transports in Fram Strait are not balanced. This
1663 implies that the computed heat and freshwater transports depend upon the choice of reference
1664 temperature and reference salinity. Rudels et al. (2008) put the reference salinity equal to the mean
1665 inflow salinity and the reference temperature equal to the mean outflow temperature. This implies
1666 that the inflow carries no freshwater and the outflow no heat.

1667 Computing the geostrophic transport through 16 hydrographic sections obtained
1668 between 1980 and 2005 along the sill in Fram Strait the total in and outflows ranged from 5 Sv to
1669 almost 15 Sv with an averaged inflow of 6 Sv and an average outflow close to 9 Sv. This is, as
1670 expected, less than the transports obtained by direct current measurements, but not alarmingly so.
1671 The mean net outflow, 2.5 Sv, 2.8 Sv if the southward flow on the Greenland shelf is included, is
1672 larger than that found from the current meter array.

1673 The transports through the other passages of the Arctic ocean are only in one
1674 direction. Inflows take place over the Barents Sea and through Bering Strait and runoff and net
1675 precipitation add freshwater. The only other outflows occur through the Canadian Arctic
1676 Archipelago. The shelf input and the Pacific inflow are cooled to freezing temperature but the part of
1677 the Barents Sea inflow that enters the deeper layers of the basins has mean temperatures around
1678 -0.5°C . We assume that the transports through the other passages are known and use the estimates
1679 provided by Dickson et al. (2007). It is then possible to determine the amount of heat entering and
1680 the amount of freshwater leaving the Arctic Ocean through Fram Strait and also how the advective
1681 heat is lost in the Arctic Ocean.

1682 The net volume transport, the mean outflow temperature and the northward heat
1683 transport referenced to the outflow temperature were computed for each section. The average

values 2.8 Sv, $T_{ref} = 0.7^{\circ}\text{C}$ and 25 TW obtained from the sections were taken to represent the transports and used to determine how the inflowing heat might be distributed in the Arctic Ocean.

Assuming that 1.2 Sv of the Barents Sea inflow with temperature -0.5°C enter the deeper layers and leave the Arctic Ocean through Fram Strait, this cold water transports $c \times (T_{ref} - (-0.5)) \times 1.2 \times 10^9 = 5.8 \text{ TW}$ into the Arctic Ocean. $c = 4000 \text{ J kg}^{-1} \text{ K}^{-1}$ is the heat capacity of sea water. The remaining part of the net outflow of 1.6 Sv is less saline and at the freezing point and exits as upper layer water transporting $c \times (T_{ref} - (-1.8)) \times 1.6 \times 10^9 = 16 \text{ TW}$ heat northward. The northward heat transport in the East Greenland Current of 22 TW then has to be balanced by outflowing warmer returning and recirculating Atlantic water. This implies that almost all 25 TW of the northward heat transport through the strait return south as warmer water in the East Greenland Current.

The remaining 3 TW could melt sea ice north of Svalbard and with an Atlantic inflow temperature of 3.1°C this implies that 0.15 Sv is cooled to freezing temperature and that the fraction $\phi = 2\alpha L(c\beta S_A)^{-1}$ (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 compiled by Dickson et al. (2007) Rudels et al. (2008) found that the freshwater fluxes into and out of the Arctic Ocean were almost in balance but the volume transports were unbalanced by 1.17 Sv. This suggests that the barotropic velocities added in Fram Strait are too small. However, since the chosen reference salinity is the mean salinity of the inflowing water, we can add a barotropic correction to the northward transports without affecting the freshwater balance. If a barotropic northward transport of 1.1 Sv is added also the volume fluxes become almost balanced. The northward heat transport, however, then increases.

The difference between the mean inflow temperature, 1.6°C , and the mean outflow temperature 0.7°C is 0.9°C , which increases the heat transport by 4 TW. The net outflow that has to be heated to 0.7°C decreases from 2.8 Sv to 1.7 Sv, implying that only 0.5 Sv of the cold surface

1710 water must be warmed to 0.7 °C, which requires 5 TW. The heat transport through Fram Strait that
1711 enters the Arctic Ocean then becomes 19 TW. If all this heat goes to ice melt by cooling the Atlantic
1712 water to the freezing point about 1 Sv of Atlantic water is converted to low salinity upper water. 7 Tw
1713 go to melting sea ice and 12 TW are lost to the atmosphere. This means that 0.021 Sv of ice is melted
1714 and the salinity of the upper layer becomes 34.3.

1715 These values are realistic and agree with other, more recent estimates of the Arctic
1716 Ocean heat and freshwater balances (e.g., Beszczynska-Möller et al., 2011, Rudels et al., 2015,
1717 Rudels, 2016, Bacon et al. 2022). It appears that the heat transport and the effects of the heat carried
1718 by the Atlantic water into and around the Arctic Ocean are less than commonly assumed. The
1719 existence of strong warm inflow pulses of Atlantic water could perhaps alter this situation and
1720 increase the effects of the Atlantic inflow. However, the persistence of the high temperatures in the
1721 Atlantic inflow pulses as they circulate around the Arctic Ocean basins (Polyakov et al, 2005, 2011)
1722 rather indicates that the heat remains in the Atlantic layer and the heat is eventually returned to the
1723 North Atlantic through Fram Strait.

1724 The discussion above is largely ad hoc and relies strongly on subjective reasoning.
1725 However, the inverse modeling of the exchanges between the Arctic Ocean and the world ocean has
1726 greatly improved as more observations from the different gateways have become available.
1727 Tsubouchi et al (2012) estimated the transports through all four passages (Davis Strait, Fram Strait,
1728 Barents Sea and Bering Strait) for August 2005 when all gateways were monitored and found an
1729 oceanic heat flux into the Arctic Ocean of 154 ± 44 TW and a freshwater export from the Arctic Ocean
1730 of 155 ± 65 mSv relative to the mean temperature $1.01^\circ\text{C} \pm 0.16^\circ\text{C}$ and the mean salinity 34.67 ± 0.02 on
1731 the total cross section. Since all passages were monitored the total heat and freshwater transports
1732 are here independent of reference temperature and salinity. However, the heat and freshwater
1733 transports through the individual straits may occasionally appear somewhat odd. The largest
1734 freshwater export takes place through Fram Strait but primarily as import of saline Atlantic water in
1735 the West Spitsbergen Current, not as export of low salinity water in the East Greenland Current.

1736 To estimate the water transformations taking place in the Arctic Ocean
1737 ~~Tsubouchi~~Tsubouchi et al. (2012, 2018), Brown et al. (2025) and Bacon et al. (2022) computed the
1738 in- and outflows in different density intervals, and the resulting imbalances ~~havehad~~ to be balanced
1739 by mixing across the density surfaces within the Arctic Ocean. The computations ~~reveal~~revealed that
1740 the density interval encompassing the Atlantic water ~~iswas~~ not balanced and of the inflowing 6.0 Sv
1741 only 3.2 Sv ~~return~~returned in the same density interval as modified Atlantic water. 1.0 Sv ~~hashad~~
1742 become diluted by freshwater and ~~returns~~returned as less dense upper layer water, while 1.8 Sv
1743 ~~havehad~~ been transformed into denser water and ~~return~~returned as intermediate and deep waters
1744 (e.g., Bacon et al. 2022). The formation of upper layer water is mainly due to sea ice melting north of
1745 Svalbard and the creation of denser water takes place by atmospheric cooling largely in the open
1746 water area in the southern Barents Sea.

1747 The observations at the gateways thus reveal transformations of the waters within the
1748 Arctic Ocean, processes that are not considered in the circulation model by Nøst and Isachsen (2003).
1749 ~~These~~The results above, however, are based mainly on observations, but Dörr et al. (2026) showed
1750 that the output from a global NEMO ocean hindcast simulation can describe many of the active
1751 processes and how they transform the waters entering the Arctic Ocean. The model was forced with
1752 atmospheric and runoff data from the DRAKKAR forcing set and was run from 1958 to 2015 and the
1753 outputs from 1979 to 2015 were examined. The computed net in- and outflows through the different
1754 passages are close to those observed, there are net outflows through Davis Strait (1.8 Sv) and
1755 through Fram Strait (3.0 Sv) and inflows over the Barents Sea (3.5 Sv) and through Bering Strait (1.3
1756 Sv).

1757 The focus is on water mass transformations and the circulation is described both in
1758 density space, showing the transformation of inflowing water into denser and less dense waters,
1759 similar to the approach used by Brown et al. (2025). However, the transformations are also shown in
1760 TS space, following the approach introduced by Gösta Walin (1982). The model results in similar
1761 water mass transformations as those deduced from observations in the Arctic Ocean and discussed

1762 earlier in the sections above. One significant difference, however, was that about 3 Sv of Atlantic
1763 water were transformed into denser water (Dörr et al. 2026), almost twice the transformation
1764 estimated by Brown et al. (2025 and Bacon et al. (2022). The main area of this transformation was
1765 [also here](#) located to the southern Barents Sea.
1766

1767 Section 10 The Greenland Sea

1768 The Nordic Seas were early identified as a possible source area for the cold, dense deep waters of the
1769 oceans. Wyville-Thomson and Carpenter found cold, dense bottom water south of the Faroes, Mohn
1770 identified a cold, almost homogenous water column southeast of Jan Mayen, and Amundsen's
1771 observations from *Gjøa* in the Greenland Sea in 1901 showed a cold homogenous dome of water
1772 extending from the bottom almost reaching the surface (Nansen, 1906). However, Nansen (1912)
1773 judged that the deep convection observed in the Greenland Sea had little global impact and instead
1774 postulated that the bulk of the ocean deep water is formed farther south, in the Irminger Sea
1775 southeast of Greenland, or in the Labrador Sea.

1776 In the 1970s this view began to change. Satellite observations showed a polynya in the
1777 Weddell Sea ice cover 1977 to 1979 and hydrographic observations confirmed that the polynya was
1778 related to a cold, less saline water column indicating recent convection (Gordon, 1978). This
1779 stimulated a search for and studies of convection, or chimneys (Killworth, 1979), and in the
1780 Greenland Sea satellite observations showed several rapid events of ice formation and subsequent
1781 disappearance of sea ice, suggesting sinking of cold brine enriched surface water, which was replaced
1782 by upwelled warmer water, melting the ice.

1783 Convection occurs as small-scale short events in winter, usually during bad weather
1784 conditions making it difficult to observe and study. In the 1970s field experiments dedicated to
1785 convection studies were initiated in known convection sites, in the Mediterranean Sea and in the
1786 Labrador Sea, and in the 1980s and 1990s most observational and theoretical work in the Greenland
1787 Sea was concentrated on convection and many new techniques were introduced. At the same time
1788 the importance of the Greenland Sea convection for the formation of dense deep water that could
1789 drive the Atlantic meridional overturning circulation, or the global thermohaline circulation as it was
1790 then commonly called, was realized, and perhaps more urgent, what were the consequences, if the
1791 convection and deep water formation should weaken. In retrospect, in 2026, the three decades
1792 between 1980 and 2010 were decades when the largest known changes in the Greenland Sea

1793 convection occurred, and when our understanding of the deep water formation in the Arctic
1794 Mediterranean deepened.

1795 In the mid-20th century the principal sign of deep convection, a homogenous water
1796 column extending from the surface to the bottom, had not been observed in the Greenland Sea and
1797 the early observations by Mohn and Amundsen were not fully trusted because the instruments used
1798 then were not of the same quality as those deployed during the 1950s. Metcalf (1955) suggested a
1799 sloping descent of the dense water that would not be captured by vertical hydrographic casts but this
1800 idea was never really accepted. The water column in the central Greenland Sea had structure that
1801 appeared to be conserved, even if its deep and bottom waters were renewed. Generally, it consisted
1802 of a low salinity upper layer above a deep temperature maximum and further down a salinity
1803 maximum below which the temperature and salinity decreased with depth until the homogenous
1804 Greenland Sea bottom water was encountered. How could this structure be explained?

1805 Carmack and Aagaard (1973) suggested that the deep and bottom water formation
1806 was driven by double-diffusive transports through diffusive interfaces. Atlantic water circulating
1807 around the Greenland Sea gyre penetrates towards the center while losing heat upward to the colder
1808 surface water but retaining its high salinity. When it finally reaches the center of the gyre, its density
1809 has become high enough to renew the deep water and perhaps sink to the bottom. The fact that the
1810 convecting water is not in direct contact with the atmosphere could also explain the low oxygen
1811 content observed in the Greenland Sea deep water (Carmack and Aagaard, 1973). The importance of
1812 double-diffusion was further explored by McDougall (1983), who found that the amount of dense
1813 water that could be created by this process corresponded well with existing estimates of deep water
1814 formation in the Greenland Sea.

1815 Deep convection areas are usually related to cyclonic circulation that forces the
1816 surface water towards the rim and allows the deeper layers at the center to rise as a dome towards
1817 the surface. A smaller volume of upper layer water must then be cooled before convection starts and
1818 the water column overturns more readily (Killworth, 1979, 1983; [Marshall and Scott, 1999](#)). The

1819 smaller volume also allows for the possibility that the convection occurs in a filling box mode, where
1820 the upper layer density increases in winter, becomes unstable and convects to the bottom, bypassing
1821 the intermediate layers (Baines and Turner, 1969). The difficulty with this explanation, in contrast to
1822 e. g. brine rejection over the shelves, is that it is not possible to accumulate dense water at the
1823 surface. A Rayleigh number based on the ocean depth would directly make the water column
1824 unstable, and even with a Rayleigh number based on the penetration depth of the heat anomaly
1825 from the surface (Howard, 1967) no large positive anomaly could be created before convection
1826 starts.

1827 This fact and satellite observations showing an interplay between the ice tongue
1828 Odden and the ice free Nordbukta, suggesting rapid formation and melting of the ice cover in the
1829 central Greenland Sea, opened for the possibility that open ocean deep convection could be driven
1830 by ice formation (Rudels, 1986a). As the water is cooled to freezing temperature, ice will form and
1831 rise to the surface, leaving more saline water droplets behind. These droplets sink through and their
1832 excess salinity diffuses into the ambient upper layer. As the droplets eventually lose their salinity it is
1833 instead the ambient upper layer that has become saline and dense with respect to the underlying
1834 water and starts to convect. The density and thickness of the unstable layer can be estimated from
1835 how long it takes for the salinity anomaly to diffuse, using molecular diffusion coefficients, into the
1836 ambient water, and how fast the parcels sink, computed from Stokes' resistance law.

1837 The unstable upper layer then sinks as entraining plumes or thermals into the deep
1838 until the plumes reach their neutral density level, where they spread out. Convection, however,
1839 continues and the upper part of the water column becomes gradually denser until its density is high
1840 enough for the convection to penetrate still deeper. The transformed upper layer then convects into
1841 the deep and becomes replaced by warmer water from below that starts to melt the ice and the
1842 convection stops. If the cooling persists, the upper layer is again cooled to freezing temperature and
1843 ice starts to form, brine is rejected, and another convection cycle starts (Rudels, 1986a: Rudels et al.,
1844 1999a)). Since it is cold water that convects, it is possible that the thermobaric effect, arising because

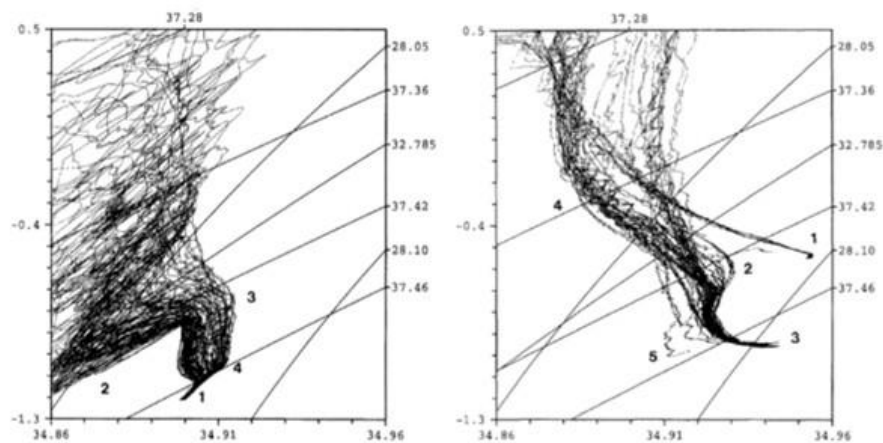
1845 cold water is more compressible than warm water, further increases the density of the plumes, if the
1846 convection reaches sufficiently deep.

1847 This scenario is highly speculative, perhaps even controversial, and field studies were
1848 initiated in the Greenland Sea to capture at least some of the involved processes (Quadfasel and
1849 Rudels, 1990). During strong cooling and ice formation a sediment trap was deployed 0.5 m below
1850 the sea surface to capture the released brine droplets. The brine was collected in sample bottles in
1851 initially deflated plastic bags. The sampling interval was 40 minutes and seven samples were
1852 collected. During the time of active ice formation three samples with salinity above the background
1853 salinity were collected. To observe the buildup and ejection of the saline top layer a 25m long drifting
1854 tube with temperature and conductivity sensors at 2.5m and 22.5m was deployed. The experiment
1855 only lasted about six hours, but a gradual increase and a sudden lowering of the salinity was
1856 observed at the upper instrument but not seen at 22.5m. This could perhaps indicate a buildup and
1857 release of a salinity anomaly (Quadfasel and Rudels, 1990; Rudels and Quadfasel, 1991.).

1858 With the advent of the CTD the resolution and accuracy of the salinity and
1859 temperature measurements increased tremendously and unexpected details were observed in the
1860 Greenland Sea water column. θS structures on stations taken in the Greenland Sea in 1993 by *FS*
1861 *Valdivia* are shown in figure 12. The densest water is the Greenland Sea bottom water concentrated
1862 in a narrow θS range with temperature and salinity decreasing towards the bottom. Above the
1863 bottom water the salinity range increases with increasing temperature and two salinity maxima can
1864 be identified. The highest salinities are found over the Greenland continental slope, along the Jan
1865 Mayen Ridge and at the Mohn-Knipovich Ridge and the lowest at the center of the Greenland Sea
1866 gyre. This implies that saline deep waters are advected from north along the Greenland slope and
1867 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).

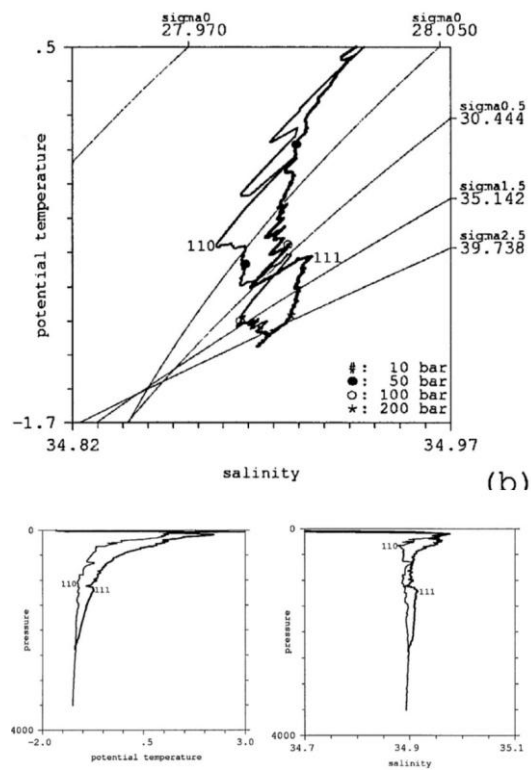
1882 The saline deep waters must ultimately arrive from the Arctic Ocean and Fig. 12 also
1883 shows θS curves from stations in the Arctic Ocean obtained by the icebreaker *Oden* in 1991. Two
1884 saline deep water modes are clearly seen. Warmer and saline deep and bottom water are observed
1885 in the Makarov Basin close to the Lomonosov Ridge. Some of this deep water crosses the sill (1800
1886 m) of the Lomonosov Ridge and enters the Amundsen Basin, where it can be identified as a salinity
1887 maximum around 1700m (Anderson et al., 1994; Björk et al., 2007). The deep and bottom waters
1888 formed in the Eurasian Basin are colder and denser but slightly less saline. The density of the
1889 Eurasian Basin Deep water corresponds well with the deeper salinity maximum observed in the
1890 Greenland Sea, while the density of the upper salinity maximum, also forming a deep temperature
1891 maximum in the central Greenland Sea, indicates the presence of the Amerasian Basin deep water.

1892 These observations show that the deep waters in the Arctic Mediterranean Sea are
1893 supplied by two sources. The open ocean convection area in the Greenland Sea ~~providing~~provides
1894 the colder, less saline endmember and the Arctic Ocean the saline, warmer endmember. The deep
1895 waters communicate through Fram Strait and depending on the regional climate, one or the other
1896 source will dominate (Aagaard et al., 1985; Rudels, 1986a).

1897 The salinity maxima are weakest in the central part of the Greenland Sea as is seen in
1898 Fig. 13 taken by *RV Lance* 1988, where an eddy exchange between the center and the rim of the
1899 Greenland Sea is captured. The exchange is isopycnal and the profiles show the doming of the
1900 density field in the central Greenland Sea, shifting the water from the rim upwards and that from the
1901 center down to deeper levels at the rim. It is thus likely that the saline Arctic Ocean deep waters that
1902 penetrate to the inner part of the Greenland Sea become diluted by the convection taking place
1903 there, largely removing their temperature and salinity anomalies.

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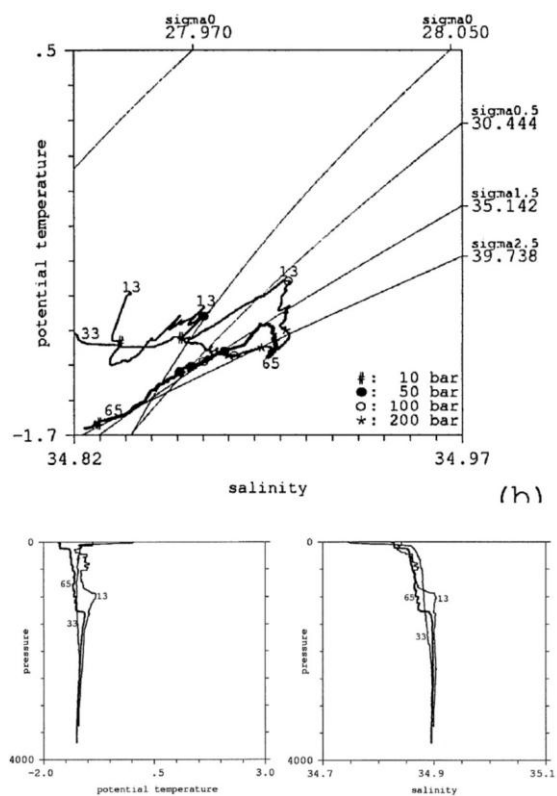


1906

1907 Figure 13. Potential temperature and salinity profiles and θ - σ curves obtained in 1988 from *RV Lance*
1908 from stations over the Greenland continental slope. The stations show an isopycnal exchange of
1909 eddies between the slope and the central Greenland Sea (adapted from Rudels et al., 1999b).
1910

1911 Figure 14 displays the potential temperature and salinity profiles and the θS curves of
1912 two stations taken in the center of the Greenland Sea in 1988, station 65 in February and station 33
1913 in June and a third station, 13, taken in 1993, all from *FS Valdivia*. The February station shows
1914 convection deeper than 1000m and the formation of a homogenous, less saline layer covered by
1915 cold, low salinity surface water and located above a more saline deep layer. In June the profiles
1916 ~~indicates~~suggest that convection has continued and reached deeper than 2000m. The salinity in the
1917 upper 1000m has increased and the salinity in the deeper part has decreased, implying that during
1918 late winter the upper layer density has increased and the two layers have been homogenized. The
1919 density of the low salinity surface layer has to increase, first by cooling to freezing temperature and if
1920 this not sufficient for convection, ice formation and brine rejection continue to increase the density
1921 to above that of the layer below, and the surface layer then convects. The density of the first layer
1922 below has then increased sufficiently for it to convect into the second deep layer below. The two
1923 layers then merge, forming a more than 2000 m deep, homogenous water column.
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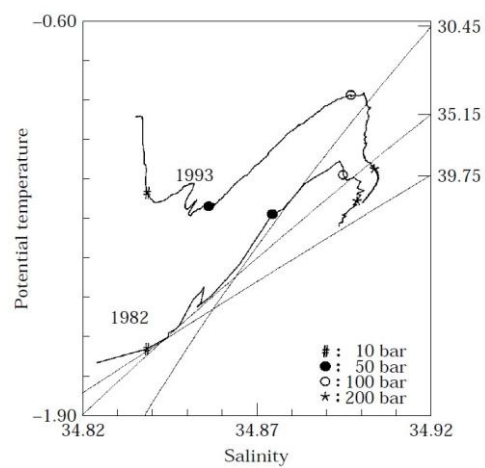
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Figure 14. Potential temperature and salinity profiles and θ - σ curves obtained by *FS Valdivia* in 1988 (65 and 33) and 1993 (13) in the central Greenland Sea, showing the evolution of the convection in during 1988 and the shallower convection observed in 1993 (adapted from Rudels et al., 1999b).

1933 Station 13 in Fig. 14 shows a warmer, more saline water column and deep coinciding
1934 temperature and salinity maxima. The convection here extends to the temperature maximum and
1935 the convected water forms a temperature minimum above the temperature maximum. The deeper
1936 salinity maximum has become more saline and the Greenland Sea bottom water salinity has
1937 increased. That the convection is shallower implies that the Arctic Ocean deep water will not be
1938 transformed in the central Greenland Sea and the deep salinity maxima become more prominent.
1939 This especially applies to the deeper maximum from the Eurasian Basin.

1940 Figure 15 shows two stations, one from 1982 taken by *RV Hudson* and one *Valdivia*
1941 station from 1993. The temperature and salinity of the deep salinity maximum have increased and
1942 also the Greenland Sea bottom water has become warmer and more saline. The salinity increase in
1943 the bottom water can only be caused by turbulent mixing from above and Meincke et al. (1997)
1944 estimated the vertical diffusion coefficient for the deep mixing in the Greenland Sea by assuming a
1945 balance between the lateral advection of Eurasian Basin deep water to the center of the gyre and the
1946 vertical mixing into the bottom water. The obtained mixing coefficient was surprisingly high, $\kappa_v \approx 3.7$
1947 $10^{-3} \text{ m}^2 \text{ s}^{-1}$. The renewal time for the deep salinity maximum was estimated to 6 years, corresponding
1948 to a through flow of 0.3 Sv. If the removal of the Greenland Sea bottom water, 0.2 Sv, is included 0.5
1949 Sv of deep water is exported from the Greenland Sea (Meincke et al., 1997). Later studies made by
1950 Somavilla et al. (2013) and Marnela et al. (2016) have indicated similar high vertical mixing
1951 coefficients.
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Figure 15. σ_t curves from two stations taken 11 years apart from *RV Hudson*, 1982 at 75°02'N 005°11'W and *FS Valdivia* 1993 at 74°00'N 004°20'W.

1959 The changes in the Greenland Sea water column continued during the 1990s and a
1960 temperature maximum started to develop at mid depth (Budéus et al., 1998), indicating that the
1961 Amerasian Basin deep water penetrating towards the center of the Greenland Sea ~~was~~ no longer ~~was~~
1962 removed or weakened by mixing with convecting cold, less saline Arctic intermediate water. The
1963 water locally formed in winter has become less dense and only the water column above the
1964 temperature maximum is ventilated by convection. The temperature maximum, by contrast, is
1965 renewed by advection, locally from the Greenland slope but ultimately from the Arctic Ocean, which
1966 gradually increases its temperature. The temperature maximum also moved downwards in the water
1967 column (Budéus and Ronski, 2009), implying that the local convection, although less dense than the
1968 water of the temperature maximum, did not stop sinking when it reached its neutral density level.
1969 Instead, its layer thickness increased, indicating that it was pressing the underlying water column
1970 downwards, by continuity forcing the bottom water towards the rim and eventually to leave the
1971 Greenland Sea.

1972 This radically changed the density structure of the central Greenland Sea. Instead of
1973 displaying a dome of dense water reaching towards the surface the density surfaces formed a bowl
1974 depressed in the center and rising towards the rim. This change implies that the doming, which was
1975 earlier considered generated by wind forcing and a necessary preconditioning for convection
1976 (Killworth, 1983; Marshall and Schott, 1999) instead is a consequence of open ocean convection.
1977 When deep convection ceases, the doming cannot be maintained and the density surfaces relax.

1978 During this phase deep reaching cold, vertically almost homogenous anticyclonic
1979 vortices were observed in the Greenland Sea (Gascard et al., 2002; Wadhams et al., 2002). Their
1980 lateral extent was 10 to 15 km, close to the local Rossby radius, which prevented them from breaking
1981 up. The vortices reached deep (>2000 m) and penetrated into the temperature maximum. The upper
1982 part of the vortices was denser but their lower part less dense than the ambient water column. The
1983 vortices were, as the rest of the Greenland Sea, capped by a low salinity layer in summer. However,
1984 the large vertical extent of the vortices made this upper layer less thick and it would more rapidly

1985 become dense in winter and convect into the vortices, recharging them with dense water. This early
1986 started and confined convection could also be a cause for the deep vertical penetration of the
1987 vortices. The fact that the vortices were less dense than the ambient water between 1500 m and
1988 2000 m made them, however, unlikely candidates for a renewal of the Greenland Sea deep and
1989 bottom water, but they stand out as intriguing features in the water column. Vortices were observed
1990 during a 10 years period but then disappeared.

1991 In 1996 a tracer release experiment to study the convection was conducted in the
1992 central Greenland Sea. A cloud of Sulphur Hexachlorid was injected in the central gyre at the density
1993 surface $\sigma_{0.5} = 30.4268$ and its spreading and mixing were monitored during the following years
1994 (Watson et al., 1999). The first winter was one with comparably strong ice formation, which likely
1995 influenced the mixing of the tracer. Outside the gyre the tracer cloud spread but little vertically but in
1996 the gyre the tracer was redistributed vertically from the surface down to 1300 m. The highest tracer
1997 concentration was reduced and the maximum was shifted slightly upward.

1998 One interpretation of these changes is that cooling or ice formation and brine
1999 rejection ~~created~~create denser water at the surface that ~~convected~~convects into the deep as
2000 entraining plumes that ~~passed~~pass through the tracer cloud and ~~redistributed~~redistribute the tracer
2001 to deeper levels. By mass continuity a compensating upward flow through the tracer cloud then
2002 ~~brought~~bring the tracer also to shallower levels. The one-dimensional convection model applied by
2003 Watson et al. (1999) agrees with this picture.

2004 With time the temperature of the cold Arctic intermediate water above the
2005 temperature maximum gradually increased and eventually it became higher than the temperature of
2006 Amerasian Basin water, which instead only was identified as a sharp salinity increase below the
2007 ventilated intermediate water. By mid 2010s the ventilated intermediate water had also become
2008 more saline than the Amerasian Basin water and its salinity maximum disappeared.

2009 This is a likely consequence of the increase in salinity and temperature of the upper
2010 layer in the Greenland Sea due to interaction with recirculating Atlantic water in the East Greenland

2011 Current, which counteracts the atmospheric cooling taking place in the central gyre (Latarius and
2012 Quadfasel, 2016). The salinity and temperature of the upper layer in the Greenland Sea are now the
2013 highest observed in the Greenland Sea water column. This has made the Greenland Sea similar to the
2014 Labrador Sea and the convection now resembles that of a gradual cooling and deepening of the
2015 upper layer in winter.

2016 The cold, less saline Greenland Sea bottom water has also disappeared and the θ_S
2017 characteristics of the bottom water are now those of the Eurasian Basin deep water (Marnela et al.,
2018 2016). The features of the water column that identified two competing sources of the deep water in
2019 the Arctic Mediterranean (Aagaard et al., 1985; Rudels, 1986a) have disappeared, and presently the
2020 Arctic Ocean source dominates and the Greenland Sea deep water is renewed by advection. If the
2021 Arctic Ocean source also ~~have~~has weakened, it has not yet been detected in the Greenland Sea
2022 because the larger reservoir present in the Arctic Ocean. The cause behind these changes is likely the
2023 warming climate that no longer creates water dense enough to ventilate the deep Greenland Sea by
2024 open ocean convection.

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2027 Section 11 Overflows.

2028 The Arctic Ocean and Greenland Sea deep waters are mostly confined to the Greenland Sea by the

2029 Jan Mayen fracture zone and the Mohn-Knipovich Ridge and contribute but little to the overflows.

2030 When the intensive studies of the Nordic Seas started in the 1970s and 1980s an outflow of

2031 Greenland Sea deep water to the Norwegian Sea was observed in the about 2200m deep channel in

2032 the Mohn Ridge just north of Jan Mayen. In the late 1990s this flow had changed direction and

2033 instead deep water from the Norwegian Sea entered the Greenland Sea, implying that the Greenland

2034 Sea deep water density had been reduced and that the deep pressure gradient had reversed

2035 (Østerhus and Gammelsrød, 1999). Since no deep water is formed in the Norwegian Sea, this

2036 indicated that water, earlier exported from the Greenland Sea, was returning. This situation probably

2037 persisted only a short period since, due to continuity, the deep water from the Arctic Ocean entering

2038 the Greenland Sea has to leave, either through gaps in the Mohn-Knipovich Ridge to the Norwegian

2039 Sea or return to the Arctic Ocean through Fram Strait.

2040 The deepest passage across the Greenland-Scotland Ridge to the North Atlantic is the

2041 Faroe Bank Channel (>840 m) and the densest deep water that enters the Norwegian Sea from the

2042 Greenland Sea must eventually exit the Arctic Mediterranean through this passage. For less dense

2043 water there is also the possibility of spillover across the 500 m deep ridge segment between Iceland

2044 and the Faroes. The Faroe Bank Channel is a continuation of the Faroe Shetland Channel and Fig. 16,

2045 showing bottom temperature and bottom salinities in the northeastern North Atlantic, reveals the

2046 narrowness of the channel and also the spillover of cold, less saline water across the Iceland-Scotland

2047 Ridge. Not all dense water that passes through the Faroe-Shetland Channel reaches the Faroe Bank

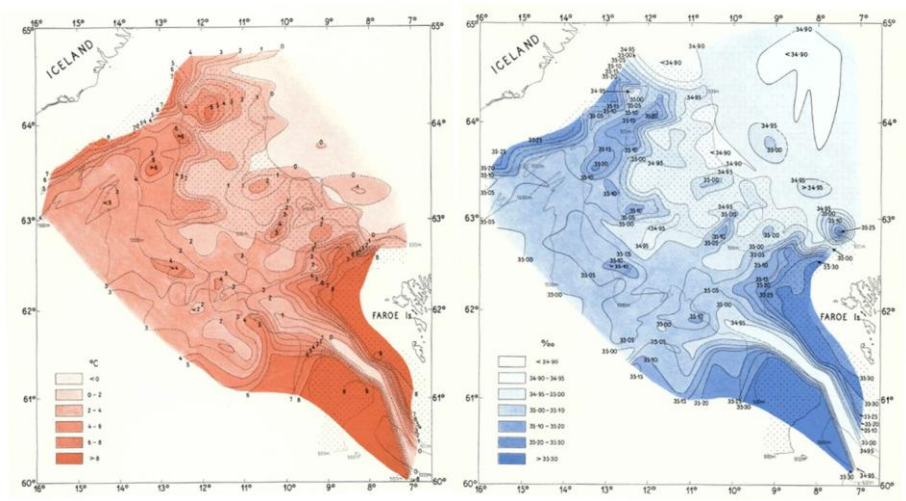
2048 Channel. Some less dense water continues south and crosses the Wyville Thomson Ridge and enters

2049 the Rockall Basin. As the name of the ridge suggests, this is where Carpenter and Wyville Thomson

2050 first observed the cold water entering the Northeast Atlantic from the north.

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2055 Figure 16. The bottom temperature (left) and salinity (right) distributions observed between Iceland
2056 and the Faroes during ICES's Overflow-60 experiment. Especially the cold and less saline water
2057 flowing through the Faroe Bank Channel is conspicuous compared to the cold water spilling over the
2058 Iceland Faroe Ridge and clearly reveal the narrowness of the channel (from Tait et al., 1967).

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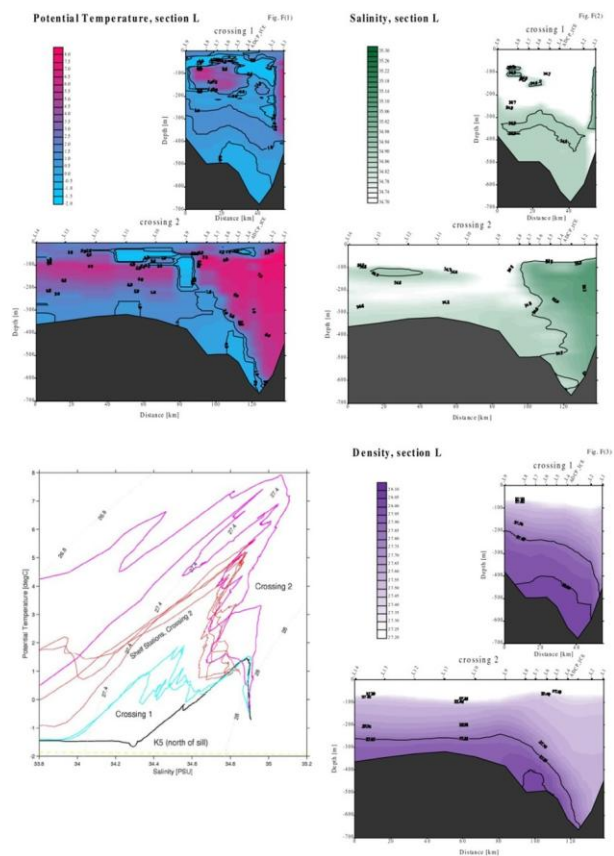
2061 Not all the overflow water from the Norwegian Sea derives from the Greenland Sea. A
2062 substantial fraction of the East Greenland Current that crosses the Jan Mayen Fracture Zone and
2063 enters the Iceland Sea continues eastward in the East Icelandic Current into the Norwegian Sea. The
2064 East Iceland Current is deep, about 600 m, and carries not only low salinity surface water but also
2065 denser waters that especially contribute to the spillover across the Iceland-Faroe Ridge. In a drifter
2066 experiment with RAFOS floats [in the Iceland Sea](#) with a drifting depth of 600m ~~in the Iceland Sea~~ all
2067 of the floats deployed east of the Kolbeinsey Ridge moved southeastward into the Norwegian Sea,
2068 and also some deployed west of Kolbeinsey Ridge entered the Norwegian Sea. However, most of the
2069 floats deployed west of the ridge either exited through Denmark Strait or grounded on the
2070 northwestern Icelandic shelf (De Jong et al., 2017).

2071 The eastern overflow entrains strongly the ambient waters and as the overflow
2072 reaches its neutral density level the overflow volume has more than doubled. The overflow water
2073 moves toward the equator but when it reaches the Gibbs Fracture Zone in the Mid-Atlantic Ridge it
2074 splits and some, maybe the major part, leaves the Iceland Basin and enters the Irminger Basin and
2075 then moves north towards Iceland. The initial low temperature and salinity characteristics of the
2076 eastern overflow has rapidly been removed and it has attained a potential temperature between 2°C
2077 and 3°C and a salinity around 34.925.

2078 The western overflow in Denmark Strait is more complicated. It is not a deep, almost
2079 stagnant reservoir that gradually drains south over a sill, but instead a region where several current
2080 branches and water masses converge. The upper part of the East Greenland Current that passes over
2081 the sill depth of the Jan Mayen Fracture Zone (1700m) comprises not only low salinity Polar water,
2082 but also Arctic Atlantic and denser intermediate waters from the Arctic Ocean, as well as warm and
2083 saline Atlantic water recirculating in Fram Strait and Arctic Intermediate waters created in and
2084 exported from the Greenland Sea. There is also the recently discovered North Icelandic Jet roughly
2085 following the 600m isobath on the northern Iceland slope westward to Denmark Strait (Jónsson and
2086 Valdimarsson, 2004). In addition, the warm and saline North Icelandic Irminger Current enters the

2087 Denmark Strait from the south. The sill depth in Denmark Strait is 640m, but the deep channel is
2088 narrow and waters from north and south contend for space and passage. Changes take place from
2089 one week to the next and Fig. 17 shows two sections along the sill taken by *RV Aranda* one week
2090 apart. On the first section waters from the north dominates, on the second water from the south.
2091 The second section extended farther towards Greenland because of favorable ice conditions (Rudels
2092 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).

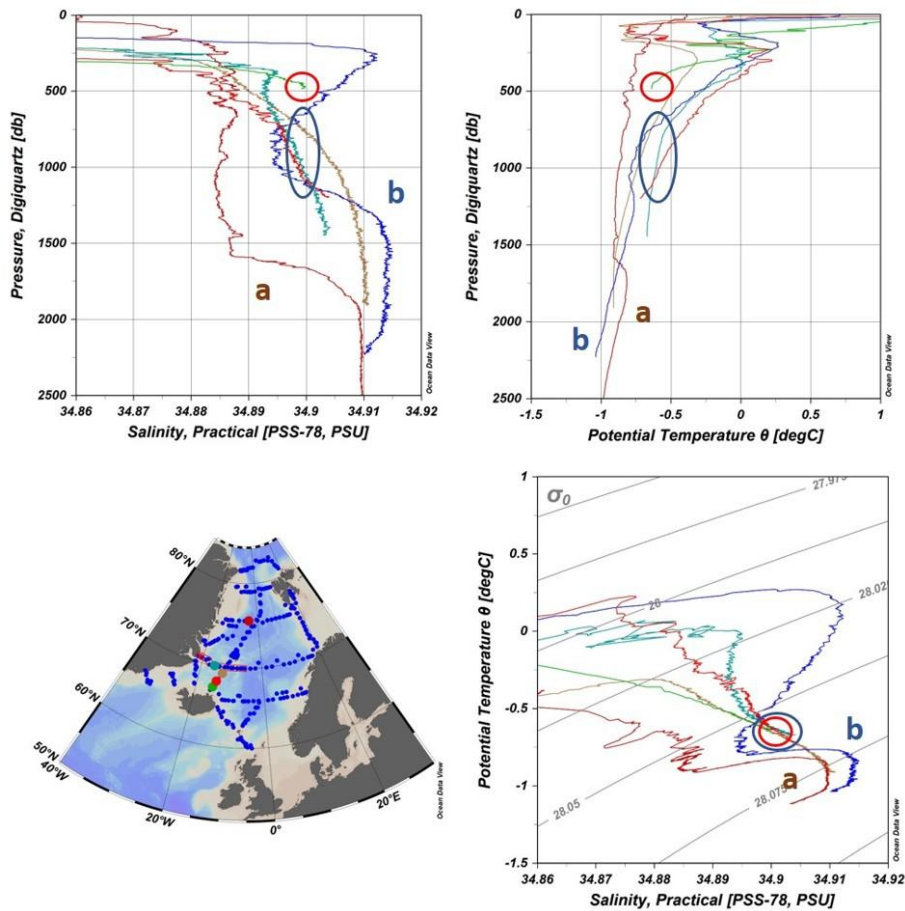
2103 Overflow water is commonly defined as water with potential density above 27.8 and
2104 the second section shows that waters with density above 27.8 are present on the shelf, which implies
2105 that even when the North Iceland Irminger Current dominates in the deep channel the overflow, or
2106 at least part of it, could instead take place over the shelf and then sink down the continental slope
2107 into the Irminger Basin. However, the East Greenland Current is not the only source of overflow
2108 water. In fact, the densest water crossing the sill is supplied by the North Icelandic Jet. The jet has
2109 been traced eastward towards its unknown source area and it is observed along the entire northern
2110 slope (Våge et al., 2011).

2111 Våge et al. (2011) suggested that the water of the North Icelandic Jet is supplied by
2112 two sources, the first would be the cooling of the North Icelandic Irminger Current as it flows
2113 eastward on the shelf. The second contribution would come from mixing with the denser water in
2114 the interior of the Iceland Basin as the cooled water returns towards Denmark Strait, further
2115 increasing its density. The problem with this explanation is that the atmospheric cooling observed in
2116 the area is not strong enough to create the observed θ_S properties of the jet (Segtman et al., 2011;
2117 Latarius and Quadfasel, 2016). Furthermore, the seasonal convection in the Iceland Sea is usually
2118 limited to the upper 200 m and only in the northern part of the Iceland Sea does the winter mixed
2119 layer occasionally attain the θ_S observed in the North Icelandic Jet (Våge et al., 2015).

2120 That the formation of dense water in the Iceland Sea in winter and its subsequent
2121 export in summer and fall was the main process supplying the Denmark Strait overflow was proposed
2122 by Swift et al. (1980) and Swift and Aagaard (1981) as an alternative to the cooling of the Atlantic
2123 water in the Norwegian Sea and its recirculation in Fram Strait and export in the East Greenland
2124 Current suggested by Worthington (1970). However, Mauritzen (1996a, 1996b) did not find any
2125 strong seasonality in the overflow, which speaks against seasonal sources in the central basins of the
2126 Nordic Seas and rather indicates a steady flow, which could be supplied by the gradually cooled
2127 Atlantic water returning to the North Atlantic through Denmark Strait.

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2128 Waters with the θ_S characteristics of the deepest overflow are present in the northern
2129 Iceland Sea and in the East Greenland Current. These θ_S properties are, however, observed at much
2130 deeper levels, 750m to 1500m, than the about 500m depth, where the core of the North Icelandic Jet
2131 is found. Moreover, it is likely that these waters are formed farther north and have been advected
2132 across the Jan Mayen Fracture Zone to the Iceland Sea. If the East Iceland Current transports a major
2133 part of the upper 600m of the water column to the Norwegian Sea, the deeper levels might, if they
2134 are advected southward, ascend to shallower depth as they approach the Iceland Continental slope
2135 and then feed the North Icelandic Jet (Fig. 18; Rudels, 2021).
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2139 Figure 18. Profiles of potential temperature and salinity and θS curves from stations in the Iceland
 2140 Sea and the Greenland Sea. The green station (red circles) found at 480 m depth at the Iceland slope
 2141 has similar θS properties as the densest overflow and also properties akin to those observed below
 2142 750 m and still deeper on stations farther north in the Iceland Sea, indicated by blue ellipses. These
 2143 properties are located at deeper levels, if a warmer Atlantic layer also is present, suggesting that it is
 2144 pushed downward by the Atlantic water. The θS curves from these stations have similar shape as the
 2145 upper Polar deep water (temperature decreasing and salinity increasing with depth), but are colder
 2146 and located lower in the θS diagram. The θS properties could be created by mixing between the
 2147 upper Polar deep water and the less saline and colder Arctic Intermediate water formed in the
 2148 Greenland Sea. However, the smooth θS curves do not indicate recent interleaving and mixing but
 2149 rather suggest long residence time and slow renewal. Stations a and b shows the difference between
 2150 the central gyre and the southern rim of the Greenland Sea. The observations are from from RV
 2151 Knorr 2002 (Figure from Rudels, 2021).

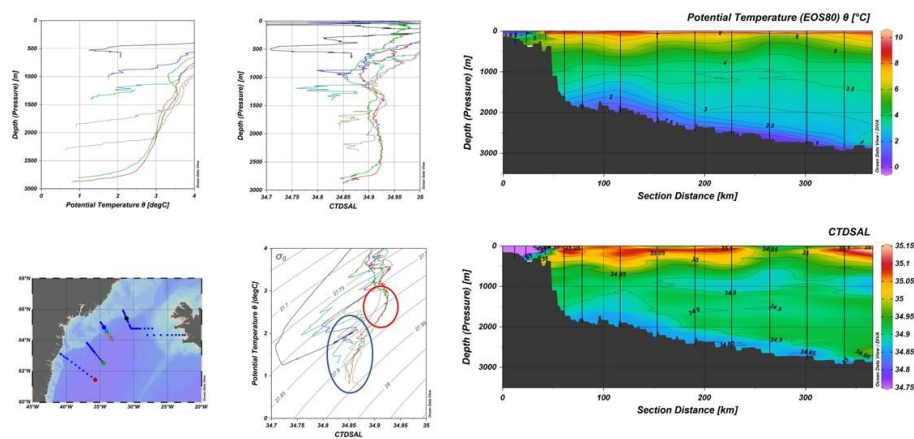
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2153 As a curiosity, the deep salinity maxima shown in Fig. 18 derive from the Jan Mayen
2154 Fracture Zone (blue station) and the central Greenland Sea (red station) and have about the same
2155 density, but the maximum at the station at the rim is located at shallower depth than the maximum
2156 at the station in the central Greenland Sea. This implies that the convection in the Greenland Sea has
2157 compressed the underlying water column, creating the deep part of a bowl, which then rises towards
2158 the rim. These stations were taken in 2002, when the Greenland Sea convection had become
2159 shallower, and it should be compared with the stations in Fig. 13 in section 10, taken in 1988, when
2160 the denser water is found at shallower levels in the center than at the rim.

2161 The overflow plume in Denmark Strait is stratified. It comprises a cold, $< -0.5^{\circ}\text{C}$,
2162 bottom layer with salinity ≈ 34.9 below a temperature and a salinity maximum of 2°C and 34.91
2163 respectively, and it is capped by a salinity minimum, a low salinity lid (Figure 19). As the plume leaves
2164 the sill its bottom temperature increases rapidly to above 0°C but remains below 1°C even when the
2165 plume has descended deeper than 3000m in the Irminger Sea, and the low salinity lid is present
2166 almost as deep. The initial rapid increase in bottom layer temperature suggests that a direct mixing
2167 between the overflow water and the northward flowing North Iceland Irminger Current could take
2168 place already at the sill in Denmark Strait. There the deepest and densest part of the overflow
2169 becomes in almost direct contact with the warm and saline Irminger Current water, and the observed
2170 increase in plume temperature in the deep Irminger Basin could result from a mixing ratio at the sill
2171 of one part of entrained water to four parts of overflow water. This is significantly less than if the
2172 entrainment would occur deeper down the slope. It is also less than the estimated entrainment for
2173 the Faroe Bank overflow.

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2178 Figure 19. Potential temperature and salinity profiles and θS curves from the Greenland slope south
2179 of Denmark Strait. The sections show the overflow as a cold, less saline bottom layer, the warm
2180 Irminger Current in the upper layer and the presence of the more saline and warmer Iceland-
2181 Scotland overflow water below 2000 m. The profiles and the θS curves reveal the downstream
2182 evolution of the overflow properties, especially the low salinity lid but also the difference between
2183 the two overflows, the Iceland-Scotland overflow indicated by red ellipse and the Denmark Strait
2184 overflow by blue ellipse in the θS diagram. The potential temperature and salinity distributions on
2185 the southernmost section are shown (data from *RV Alexander von Humboldt* 2004) (Figure from
2186 Rudels, 2021).

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2189 The temperature minimum and the low salinity lid, by contrast, must derive from the
2190 Greenland shelf, where waters of the observed properties are found. The fact that especially the low
2191 salinity lid survives as the plume descends deeper than 2000m also indicates that the entrainment of
2192 ambient water through the upper boundary of the plume is weak. However, the low salinity lid
2193 suggests that if the less dense water can remain in contact with the densest part of the plume in the
2194 early stages, when it starts to descend the slope, it will remain attached to the plume and sink into
2195 the deep. One factor that can facilitate the descend of the cold low salinity upper part of the plume is
2196 the thermobaric effect, which increases the relative density of colder water as the pressure
2197 increases.

2198 As the Denmark Strait plume descends it eventually sinks through the overflow water from the Faroe
2199 Bank channel that shows up as a salinity maximum above the colder, less saline Denmark Strait
2200 overflow. Both overflows then continue to the Labrador Sea.

2201 If the dense water that flows southward on the Greenland shelf reaches the shelf
2202 break farther south, where the shelf narrows, the entrainment appears to be stronger and the
2203 plumes will remain on the upper part of the slope and not connect to the deeper overflow. This
2204 ~~appears to could~~ be the case for the Greenland spill jet, identified by the density surfaces rising
2205 towards the shelf, which indicates velocities increasing with depth (Pickart et al., 2001). It flows as a
2206 separate core above the deep overflow plume towards Cape Farewell and eventually the density
2207 surfaces flatten and the spill jet disappears (Pickart et al., 2001). ~~As the Denmark Strait plume~~
2208 ~~descends it sinks through the overflow water from the Faroe Bank channel that shows up as a salinity~~
2209 ~~maximum above the colder, less saline Denmark Strait overflow. Both overflows then continue to the~~
2210 ~~Labrador Sea.~~

2211 The overflow in Denmark Strait is strong and varies on time scales of a few days (two
2212 to five days) The strong overflow events were noticed by Worthington (1969) as the first direct
2213 current meter arrays was recovered. These strong overflow events had periods slightly less than a
2214 week. However, later observations (~~Van~~Von Appen et al., 2017) have shown that two different kinds

2215 of strong overflow events are present. The boluses are related to overflow events where the
2216 thickness of the overflow plume increases drastically, the temperature of the boluses decreases, but
2217 no significant increase in the overflow velocity is observed. The second overflow anomaly ~~Van~~Von
2218 Appen et al. (2017) denoted as pulses. The pulses show an increase the overflow velocity but the
2219 thickness of the overflow layer decreases. The temperature reduction is less (-0.1 °C) than that
2220 observed for the boluses (-0.4 °C) and the more rapid flow in the pulses compensates for the thinner
2221 overflow plume and the overflow strength is about the same for the boluses and the pulses.

2222 The difference of height of the overflow plumes first suggests a supercritical flow for
2223 the boluses and a subcritical flow for the pulses, but nature is most likely more devious than this. The
2224 flow is rotating and the thickening of the boluses is associated with cyclonic rotation and the thinning
2225 of the pulses relates to anticyclonic rotation: (Von Appen et al., 2017). The most interesting question
2226 is, however, if the boluses and pulses are associated with different water masses. The almost similar
2227 temperature of the overflow water in the two cases suggests that the main sources are in the central
2228 gyres, not in the East Greenland Current, perhaps the Greenland Sea rather than the Iceland Sea.

2229 The overflows contribute together with the dense water formed by open ocean
2230 convection in the Labrador Sea to the North Atlantic deep water and to the Atlantic Overturning
2231 Circulation. The overturning circulation has been monitored since 2004 by the RAPID-MOCHA
2232 (Meridional Overturning Circulation and Heat-flux Array) array deployed at 26°N. One of the
2233 drawbacks with this array is that it is located too far south to separate the overturning circulation,
2234 sinking into the deep, from the horizontal recirculation of cooled upper water in the subtropical gyre.
2235 An additional array, OSNAP (Overturning in the Subpolar North Atlantic Program), located north of
2236 the subtropical gyre has therefore been proposed and accepted, and the first deployment was in
2237 2014. The array extends from Scotland to Greenland (OSNAP east) and from Greenland to Labrador
2238 (OSNAP west). The first deployment ended in 2016 and when the results from the first 20 months
2239 were published (Lozier et al., 2019) the title “A sea change in our view of overturning in the subpolar
2240 North Atlantic” suggests that the results, at least for some of the researchers involved, have come as

2241 a surprise. The convection and dense water formation in the Labrador Sea contributes, as estimated
2242 by the OSNAP west array, but little, 2.1 ± 0.3 Sv, to the overturning circulation. By contrast the
2243 transformation to denser water measured by the OSNAP east array was 15.6 ± 0.8 Sv, almost an
2244 order of magnitude larger. Since some recirculation takes place, the measured total overturning is
2245 somewhat less, 14.9 ± 0.9 Sv.

2246 For those who have been working in the Arctic Mediterranean Sea and on the
2247 exchanges across the Greenland-Scotland Ridge these numbers are not surprising and should also be
2248 familiar to those remembering L.V. Worthington's 1976 monograph on the North Atlantic circulation.
2249 The observations made during the NACLIM project (North Atlantic CLIMate) indicated an overflow of
2250 5.8 Sv (Østerhus et al., 2019). Assuming a doubling of the volume due to entrainment into the
2251 plumes and some contribution from the spill jet the numbers almost match.

2252 The small volume of Atlantic deep water formed in the Labrador Sea comes as a
2253 surprise though. It suggests that seasonal deep convection is not a process that creates and drives an
2254 overturning circulation. It rather forms MODE waters than deep water (McCartney, 1977; McCartney
2255 and Talley, 1982). To create the dense water that supplies the overturning circulation, a sill and an
2256 isolated basin might be required. This would inject water with substantially higher density contrast
2257 that can sink down and entrain ambient water, creating a significant volume of deep water with
2258 distinct properties. The fact that also the much weaker Mediterranean overflow shows up in and
2259 characterizes the Atlantic water column over a large region suggests that this could be the case.

2260 Observations from the RAPID array indicates a reduction of the overturning circulation
2261 by 2.5 Sv in 2009. The weaker overturning has persisted but no further reduction has been observed
2262 (Bryden et al., 2020). The reduction has, surprisingly, taken place in the denser part of the
2263 overturning circulation, which is supplied by the overflow from the Arctic Mediterranean. During the
2264 same time the EU projects THOR (ThermoHaline Overturning at Risk) and NACLIM were running and
2265 no reduction was observed, not in the overflows, nor in the northward inflows of Atlantic water to
2266 the Arctic Mediterranean.

2267 The RAPID array showed that reduction of the overturning circulation was
2268 compensated by an increase in the recirculation in the subtropical gyre by 2.5 Sv, about 15% of the
2269 overturning circulation. This implies that less warm Atlantic water is transported by the North
2270 Atlantic drift to the subpolar gyre, which should lead to a cooling of the northeast Atlantic.
2271

2272 Section 12: Polar outflows and the Canadian Arctic Archipelago and Baffin Bay

2273 The water leaving the Arctic Mediterranean Sea consists not only of dense overflow water but also of
2274 cold, less saline and less dense polar water that interacts with the upper part of the subpolar North
2275 Atlantic. There are two main, interacting, outflows, the outflow of Polar Water through Fram Strait in
2276 the East Greenland Current that follows the Greenland shelf and coast southward through Denmark
2277 Strait to Cape Farewell, and the polar water that flows through the narrow and shallow passages in
2278 the Canadian Arctic Archipelago to Baffin Bay and then continues through Davis Strait to the
2279 Labrador Sea. These two flows interact. A part of the East Greenland Current turns north after it has
2280 rounded Cape Farewell and forms the West Greenland Current, which enters Baffin Bay from the
2281 south. The upper, polar, part of West Greenland Current mixes with the outflows from Nares Strait,
2282 Jones Sound and Lancaster Sound and the mixing product is transformed by ice formation during
2283 winter into a distinct, about 200 m deep upper layer with temperature of -1.6°C and salinity of 33.7
2284 (Fig. 20).

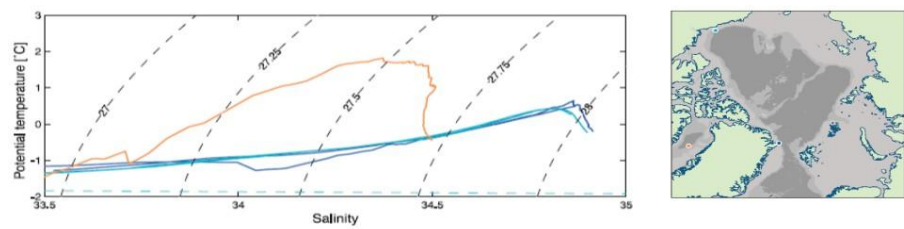
2285

2295 Below the low salinity upper water, a warm, 3-4°C, but rather fresh, 34.5-34.6, Atlantic
2296 layer is present. The Atlantic water enters Baffin Bay through Davis Strait in the West Greenland
2297 Current and the low salinity indicates that it was originally a part of the Irminger Current that has
2298 become diluted by cold, low salinity water from the East Greenland Current south of Denmark Strait.

2299 The bottom water in Baffin Bay is again colder, about –0.45°C with salinity 34.45. The
2300 origin of the Baffin Bay bottom water has been debated, Sverdrup et al. (1942) considered it to be a
2301 mixture between Labrador Sea water and Baffin Bay surface water, and local formation by
2302 convection was suggested by Bourke and Paquette (1991). Bailey (1956) found TS characteristics at
2303 sill depth north of Nares Strait similar to those of the Baffin Bay bottom water and proposed that
2304 bottom water entered from the Arctic Ocean. Figure 21 shows the TS curves from one station in
2305 Baffin Bay and two stations from the Arctic Ocean, one from the continental slope in the Canada
2306 Basin and the other from the slope northeast of Greenland, indicating that the Baffin Bay bottom
2307 water could derive from the Arctic Ocean.

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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).

2318 The TS curves also show the warm but less saline characteristics of the Baffin Bay
2319 Atlantic water compared to the Arctic Atlantic water to the north. They also, interestingly, indicate
2320 that the entrainment of ambient water into the deep inflow must be small. If the inflow entrains
2321 much, it will become warmer, which reduces its density. This, in turn, requires a denser initial inflow.
2322 However, that would also mean a warmer inflow, which would lead to different TS characteristics in
2323 the Baffin Bay bottom water.

2324 To determine the export of low salinity polar water by direct current measurements is
2325 difficult. Drifting ice and bad weather conditions complicate not only the deployment and recovery
2326 but also the survival of the moorings. Since one primary object is to determine the freshwater fluxes
2327 that may influence the convection in the subpolar gyre, it is essential to obtain measurements close
2328 to the surface and the ice, where most of freshwater is located. In spite of these difficulties
2329 measurements of volume and freshwater fluxes have been made, at least for shorter periods, in all
2330 the major passages; Lancaster Sound (Peterson et al., 2012), Jones Sound (Cardigan Strait and Hells
2331 Gate) (Melling et al., 2008), and Nares Strait (Münchow et al., 2008) in the Canadian Arctic
2332 Archipelago, in Davis Strait (Curry et al., 2011, 2014), and on the shelf area east of Greenland (De
2333 Steur et al., 2009; De Steur et al., 2016). The techniques used in the different passages are distinct
2334 and adapted ~~for~~to the local challenges such as the presence of sea ice and the closeness to the
2335 magnetic North Pole that complicates the direction measurements. Details of these observations are
2336 given in Melling et al. (2008) and in Rudels (2021).

2337 The difficulty to perform field observation, especially for monitoring purpose, makes it
2338 tempting to use existing simple models to estimate the transports of Polar water through the
2339 different passages. One approach follows Werenskiöld (1935), who derived an expression for the
2340 geostrophic transport M in a coastal current based only on the depth of the upper layer at the coast,
2341 H_1 , and the density difference $(\rho_2 - \rho_1)$ between the upper layer and the lower layer at rest, which
2342 extends to the surface one Rossby Radius away from the coast.

2343

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

2345

2346 The density in cold Arctic waters is mainly controlled by salinity and can be written as $\rho = \rho_r(1 + \beta S)$,
 2347 where ρ_r is the reference density and β the coefficient of salt contraction. Taking the density of the
 2348 lower layer as the reference density the reduced gravity ρ' becomes $g\beta(S_2 - S_1)$.

2349 This represent the upper circulation and the amount of lower layer Atlantic water that
 2350 enters the Arctic Ocean and is entrained into the upper layer. Stigebrandt (1981) modelled the
 2351 entrainment with a Kato-Phillips mixing model and noticed that if the openings are wider than a
 2352 Rossby radius they all transport with the same rate.

2353 There are three straits that are wider than the Rossby Radius and can transport at full
 2354 capacity, Fram Strait, Lancaster Sound and Nares Strait. The total transport through these passages
 2355 comprises the liquid freshwater flux F added to the water column by rivers ~~F~~ and by the low salinity
 2356 Pacific inflow L and the entrained lower layer water M_2 . The Pacific inflow is here separated into
 2357 freshwater and lower layer water fluxes. This gives:

2358

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

2360

2361 and the thickness of the freshwater m in the upper layer becomes:

2362

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

2364

2365 The salinity of the upper layer can ~~then~~ be written as:

2366

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

2368

2369 Introducing this in the expression for the outflow gives

2370

$$2371 \quad (M_2 + F) = \left(\frac{3g\beta FS_2}{2f} \right)^{\frac{1}{2}} H_1 \quad (8)$$

2372

2373 and m becomes:

2374

$$2375 \quad m = \frac{S_2(M_2+F) - (M_2S_2)}{S_2(M_2+F)} H_1 = \frac{FH_1}{(M_2+F)} = \left(\frac{2fF}{3g\beta S_2} \right)^{\frac{1}{2}} \quad (9)$$

2376

2377 The Rossby Radius Ro is independent of the entrainment and only depends on the freshwater input:

2378

$$2379 \quad Ro = \frac{[g\beta(S_2-S_1)H_1]^{\frac{1}{2}}}{f} = \left(\frac{2g\beta S_2 F}{3f^3} \right)^{\frac{1}{4}} \quad (10)$$

2380

2381 The Rossby radius is about 10 km, which confirms that both Lancaster Sound and Nares Strait are

2382 wide enough to transport with full capacity.

2383

2384 If we assume that the depth H_1 is the same in all channels and equal to the depth of

2385 the upper layer in the Arctic Ocean, the liquid freshwater storage in the Arctic Ocean can be

2386 estimated. Introducing $g = 9.83 \text{ ms}^{-2}$, $f = 1.4 \times 10^{-4} \text{ s}^{-1}$, $\beta = 8 \times 10^{-4}$, $F = 0.20 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ and setting S_2 equal

2387 to 35 and the salinity of sea ice to zero the liquid freshwater storage becomes 8.24 m. The area of

2388 the Arctic Ocean is about $10 \times 10^{12} \text{ m}^2$ and with a freshwater input/export of 0.2 Sv the residence time

2389 becomes 13 years, or half of this if only the deep part of the Arctic Ocean is considered. The

2390 freshwater storage increases as $F^{1/2}$ but its residence time decreases as $F^{-1/2}$ (Rudels, 2010).

2391 The thickness of the ice cover was in the 1960s estimated to 3 m, about one third of

2392 the liquid freshwater storage, but had decreased by about 40% by the 1990s (Rothrock et al., 1999).

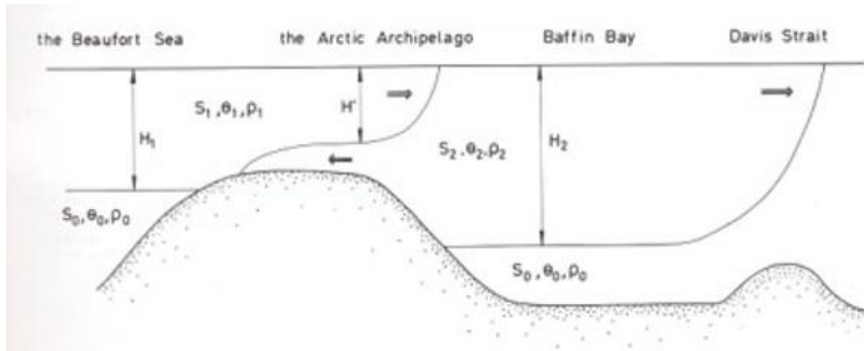
2393 ~~and if~~. If the ice export $F_i = 0.909$ Sv, remains constant, ~~is the mean~~ residence time of the ice cover is
2394 between 2.6 and 5.3 years depending upon the area considered. If only a seasonal ice cover were
2395 present and no sea ice were exported the liquid freshwater storage would increase to a little above
2396 10 m (Rudels, 2010). The liquid freshwater in the Arctic Ocean is, however, not a lid lying on top of a
2397 deeper water mass but is distributed over a large, spatially varying, depth range in the Arctic Ocean
2398 water column. The conditions in the three passages are also not equal. Especially the characteristics
2399 of the second layer differ between the passages, which affects the transports of both volume and
2400 freshwater.

2401 The second passageway from the Arctic Ocean to the North Atlantic, the Canadian
2402 Arctic Archipelago and Baffin Bay, offers additional challenges but also other opportunities to
2403 theoretically estimate the transports. The geostrophic transports depend upon the number of
2404 openings (Stigebrandt, 1981) and two passages, Lancaster Sound and Nares Strait, transport Polar
2405 water from the Arctic Ocean to Baffin Bay. In addition, the West Greenland Current carries water,
2406 ultimately from Fram Strait and the Arctic Ocean, into Baffin Bay from the south. These inflows have
2407 to exit through one opening, Davis Strait, to the Labrador Sea.

2408 The outflow from the Arctic Ocean is ultimately due to the higher sea level in the
2409 Arctic Ocean, which implies that the upper layer in Baffin Bay must be denser than that in the Arctic
2410 Ocean and its depth should be greater than the depth of inflows to Baffin Bay. There will be a
2411 pressure reversal in depth in Baffin Bay, below which the pressure gradient forces the Baffin Bay
2412 water to enter the inflow channels. The water from Baffin Bay does not enter the Arctic Ocean but it
2413 forms a lower boundary below the inflows that is taken as the depth of the transports to Baffin Bay.
2414 The depth and density of the Baffin Bay upper layer determine the geostrophic transport out of
2415 Baffin Bay, which should balance the northern inflows. The Arctic Ocean is connected to the deeper
2416 layers in Baffin Bay through Fram Strait and across the sills of the Greenland-Scotland Ridge and
2417 Davis Strait, removing any deep pressure gradient (Figure 22).

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2422 Figure 22. Idealized three reservoir system representing the Beaufort Sea, the Canadian Arctic
2423 Archipelago, Baffin Bay, Davis Strait and Labrador Sea system (from Rudels, 1986b).

2424

2425

2426 Assuming that the temperature and salinity of the Baffin Bay upper layer are -1.6°C
2427 and 33.7, the upper layer depth in Baffin Bay can be found, and the geostrophic transports through
2428 the system can be determined as functions of the Polar water salinity or density (Rudels, 1986b). The
2429 transports decrease with increasing salinity of the Polar water, since the density difference between
2430 the layers is then ~~is~~ reduced.

2431 However, the characteristics of the Baffin Bay upper layer are not only due to the
2432 mixing between the entering waters but also influenced by the cooling and ice formation taking place
2433 in winter. The cooling and the ice formation only depend on the atmospheric forcing and are
2434 independent of the inflows, and a higher salinity of the entering Polar water would require an
2435 increased transport into Baffin Bay to reproduce the observed salinity of the Baffin Bay upper water.

2436 Since one estimate gives a transport decreasing with increasing salinity and the other
2437 an increasing transport with increasing salinity, a salinity must exist for which these transports are
2438 equal. This salinity is then set as the salinity of the Polar water leaving the Arctic Ocean. The Polar
2439 outflow, the freshwater export, and the ice export through the Canadian Arctic Archipelago-Baffin
2440 Bay system can then be determined (Rudels, 1986b). However, the outflow estimate obtained by this
2441 approach, 0.7 Sv, was lower than the 2 Sv then commonly cited. One serious weakness of this
2442 approach is that it requires a residence time for the Polar water in Baffin Bay of at least one year for
2443 the transformation into the Baffin Bay upper layer to take place. Especially the inflow through
2444 Lancaster Sound that enters close to Davis Strait may pass through Baffin Bay into the Labrador Sea
2445 in a time shorter than one year.

2446 The expected change into a warmer climate has brought melting of the Greenland ice
2447 cap into focus. Intensified melting has been observed in the Jacobhavn isbrae in Baffin Bay, but also
2448 the low salinity East Greenland Coastal Current (Sutherland and Pickart, 2007), flowing on the
2449 Greenland shelf inshore of the East Greenland Current, indicates that freshwater is supplied from
2450 Greenland as runoff and as ice melt. This freshwater input, either directly to Baffin Bay and/or
2451 advected through Davis Strait into Baffin Bay, will lower the salinity of the upper layer in Baffin Bay.

2452 The density contrast between the Polar water and the Baffin Bay upper layer then
2453 diminishes and the transport through the Archipelago becomes smaller. The expected increase in
2454 mass and freshwater fluxes from the Arctic Ocean to the North Atlantic in a warmer climate, might
2455 then not take place through the Canadian Arctic Archipelago. Instead, much of this increased outflow
2456 could be channeled through Fram Strait, which might affect the dense water formation and
2457 convection both in the Greenland Sea and in the Irminger and Labrador seas (Rudels, 2011).

2458 It should be noted that in Rudels (2011) the ice formation and ice export are
2459 computed from the mass and salt balances in Baffin Bay using the postulated salinity of the Baffin
2460 Bay upper layer. It is then not independent of the inflow salinities. This in contrast to the approach in
2461 Rudels (1986b), where the ice formation was considered due to the atmospheric heat loss and
2462 unrelated to the salinities of the transports. It could then be used, together with the salinity of the
2463 upper layer, to determine the salinity of the Polar water entering Baffin Bay. The author appears, 25
2464 years later, to have forgotten this aspect.

2465 The Baffin Bay water column comprises not only transformed Polar water but also
2466 Atlantic water advected from the south and cold deep and bottom waters originating in the Arctic
2467 Ocean. These waters must eventually leave Baffin Bay, which requires a deep circulation in Baffin Bay
2468 that has not, so far, been considered. Rudels (2011) tried to theoretically estimate this deep
2469 circulation and how it relates to the upper layer exchanges by assuming that the inflows of upper
2470 water through the northern channels and in the West Greenland Current, ~~nowhere~~ considered as
2471 part of the Polar inflow, are shallower than the depth of the pressure reversal.

2472 This allows for inflows from the deeper layers to Baffin Bay. Furthermore, the pressure
2473 reversal between the Arctic Ocean and Baffin Bay does not occur in Baffin Bay but in the Labrador
2474 Sea, below the sill depth in Davis Strait. The pressure head due to the higher sea level in the Arctic
2475 Ocean is thus not removed in Baffin Bay and it forces not only a deep inflow in the north but also a
2476 deep outflow to the Labrador Sea. The pressure reversal in the West Greenland Current, by contrast,
2477 occurs above the sill depth and the deep Atlantic inflow recirculates in Baffin Bay and then exits

2478 through Davis strait together with, but above, the outflow of the deep waters from the Arctic Ocean
2479 (Rudels, 2011).

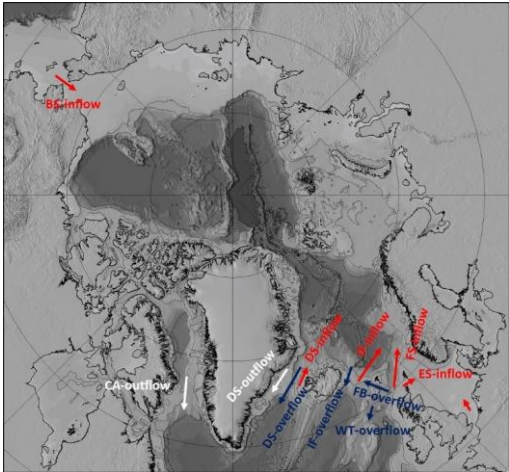
2480 The deep inflows decrease with increasing salinity of the Polar water, but the deep
2481 outflow through Davis Strait increases with the salinity of the Polar water. This implies that there is a
2482 salinity, 33.2 in this case, of the Polar water for which the deep exchanges in Baffin Bay are balanced.
2483 The deep outflow from the Arctic Ocean, taking place in the deepest channel, Nares Strait is 0.62 Sv,
2484 which is almost as large as the Nares Strait upper layer outflow of 0.77 Sv. The added barotropic
2485 pressure gradient also increases the outflow of Polar water from the Arctic Ocean to the North
2486 Atlantic and the total outflow through the Canadian Arctic Archipelago- Baffin Bay system becomes
2487 about 2.2 Sv, somewhat larger than the about 1.5 Sv presently determined from observations (e.g.,
2488 Curry et al., 2014). For details of this theoretical approach the reader is referred to Rudels, (1986b,
2489 2011).
2490

2491 Section 13. The Subpolar gyre

2492 All waters that circulate around the Arctic Mediterranean eventually flow south, entering the
2493 subpolar gyre in the North Atlantic. For most waters it is a return to where they started ~~from~~ as
2494 warm, saline Atlantic inflows crossing the Greenland-Scotland Ridge. Only the Pacific inflow ~~is~~ and the
2495 freshwater input are added to the water entering the subpolar gyre from the north. We may remark
2496 here that in spite of the Pacific inflow ~~bringing~~ brings low salinity water to the Arctic Mediterranean,
2497 it constitutes a salt input to the Atlantic Ocean that eventually crosses the equator and is exported to
2498 the Southern Ocean.

2499 The Atlantic water entering the Arctic Mediterranean becomes transformed in many
2500 ways by several different processes but, essentially, it is separated into denser and less dense waters
2501 by cooling and by receiving or rejecting freshwater, mainly through sea ice melting or sea ice
2502 formation. Waters that return to the North Atlantic either as dense overflows or buoyant outflows.
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Figure 23: Map showing the volume balance of the Arctic Mediterranean Sea. BS: Bering Strait inflow, DS: Denmark Strait inflow, IF: Iceland-Faroe inflow, FS: Faroe-Shetland inflow, ES: European shelf inflow, CA: Canadian Arctic Archipelago outflow, DS: Denmark Strait outflow, DS: Denmark Strait overflow, IF: Iceland Faroe overflow, FB: Faroe Bank overflow, WT: Wyville Thomson overflow (adapted from Østerhus et al., 2019).

2514 Østerhus et al. (2019) compiled the measured volume transports through all different
 2515 control sections to find if the volume fluxes balance (Figure 23 and Table 1). They found an inflow of
 2516 9.1 Sv and an outflow of 9.5 Sv, the exchange being almost in balance. However, an imbalance of 0.1
 2517 Sv would in one year lead to a sea level change in the Arctic Mediterranean of 0.2 m. This is more
 2518 than that observed and indicates that the direct current measurements may still be improved.
 2519
 2520 Table 1: Inflows, outflows and overflows in the different passages of the Arctic Mediterranean (based
 2521 on Østerhus et al. 2019).
 2522

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	

2523
 2524 However, what do not balance are the freshwater transports. Haine et al. (2015)
 2525 estimated the freshwater storage in the Arctic Mediterranean in two periods, between 1980 and
 2526 2000 and between 2000 and 2010 (Table 2) and discussed the reason for the changes. What is clear,
 2527 is that the adjustment time for the baroclinic pressure field is much longer than that for the
 2528 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 (km³)		
Liquid freshwater	93000	101000
Beaufort Gyre	18500	23500
Average sea ice cover	17800	14300
Total freshwater volume	110800	115300
Freshwater fluxes (km³a⁻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 (km³a⁻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

2534

2535

Of the Atlantic water about 6 Sv are transformed to dense overflow water while

2536

slightly more than 2 Sv become less dense water that together with the Pacific water make up the

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outflow waters (Østerhus et al, 2019). ~~The~~[This](#) is slightly more than the 1 Sv of less dense Polar water

2538

and the overflow of 1.8 Sv plus 3.2 Sv of overflow water estimated by Brown et al. (2025) and Bacon

2539

et al, (2022), assuming that the Atlantic water transformed in the Nordic Seas is dense enough to

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become overflow water.

2541 The overflow water sinks into the deep and eventually joins the southward moving
2542 deep western boundary current, but the outflow water circulates in the upper part of the subpolar
2543 gyre and will influence the upstream conditions of the inflows to the Arctic Mediterranean.

2544 The most conspicuous northern input to the subpolar gyre was the “Great Salinity
2545 Anomaly” observed in the late 1960s and early 1970s (Dickson et al., 1988). The outflow of low
2546 salinity water and ice through Denmark Strait strongly reduced the salinity in the gyre and the
2547 anomaly could be traced around the subpolar gyre and then into the Nordic Seas at least as far as
2548 Fram Strait (Dickson et al., 1988).

2549 Up to the mid-1990s the surface waters in the subpolar gyre were colder and less
2550 saline than their average temperature and the same conditions prevailed in the Nordic Seas and the
2551 Barents Sea. In the late 1980s to the mid-1990s the North Atlantic Oscillation (NAO) index, the
2552 normalized air pressure difference between the Azores and Iceland) was high and positive, and the
2553 cold continental air that dominated over the Labrador Sea led to a strong and deep convection and
2554 the created Labrador Sea water was exceptionally cold and fresh (Lazier et al., 2002). The cooling of
2555 the Labrador Sea ~~increased~~accelerated the circulation of the subpolar gyre, which extended far to
2556 the east and partly prevented warmer water from the subtropical gyre to enter the Nordic Seas. The
2557 high NAO index was also manifested in the changes in the Barents Sea and in the Arctic Ocean, where
2558 the river runoff was diverted farther to the east, from the Amundsen Basin to the Makarov Basin.

2559 As the NAO index eventually became lower the subpolar gyre relaxed and retreated,
2560 leaving a freer passage for the waters of the North Atlantic Drift to reach the Greenland-Scotland
2561 Ridge and enter the Nordic Sea (Hátún et al, 2005; Häkkinen et al., 2008). After this a period of
2562 almost 15 year of warm, saline surface water was observed in the subpolar gyre and in the Nordic
2563 Seas and the Barents Sea.

2564 The North Atlantic drift splits as it approaches the European continent and one part
2565 turns south and remains in the subtropical gyre. The simpleminded picture would be that it is the
2566 southernmost branch that turns south, but observations from floats launched in the upper 15 m

2567 indicated that most of the floats turned south and instead the deeper layers, around 300 m and
2568 deeper, enter the subpolar gyre (Brambilla and Talley, 2006; Burkholder and Lozier, 2011). On its
2569 path towards the Nordic Seas the water from the subtropical gyre is also cooled and freshened by
2570 mixing with the low salinity upper layers of the subpolar gyre flowing eastward from the Labrador
2571 Sea (Burkholder and Lozier, 2011).

2572 In 2009 the RAPID Array at 26 °N showed an increase of 2.5 Sv in the southward
2573 recirculation of the North Atlantic Drift (Smeed et al., 2014). This would reduce the warm water
2574 inflow to the subpolar gyre and less warm water would enter the Norwegian Sea. This change
2575 probably ended or at least contributed to the end of the 20 years period of warm surface water in
2576 the subpolar gyre (Gonzales Pola et al., 2023).

2577 In 2012 a strong freshening of the subpolar gyre was observed, the strongest in 120
2578 years (Holliday et al., 2020). This freshening appears to be different from the one associated with the
2579 Great Salinity Anomaly in that it originated in a rerouting of the circulation of the low salinity
2580 Labrador Current. Instead of moving south around Newfoundland, it was detached from the
2581 continental slope and joined the subpolar gyre circulation eastward. This low salinity signal has
2582 continued northward, into the Barents Sea and to Fram Strait (Gonzales Pola et al., 2023). If this
2583 freshening, taking place at least in the surface water, will imply a stronger stability of the Arctic
2584 Ocean water column and perhaps limit the vertical mixing in the Arctic Ocean and possibly affect the
2585 ice-ocean interactions, remains to be seen.

2586 The atmospheric circulation patterns and the wind fields thus strongly influence the
2587 circulation into and out of the Arctic Mediterranean Sea, but are they also the main causes for the
2588 overturning circulation? The overturning requires that the northward flowing water becomes denser
2589 and sinks into the deep. Moreover, it also appears to require a sill that allows the convection to build
2590 up a reservoir of denser water and create a density difference between the deep waters on the two
2591 sides of the ridge. As the dense water crosses the ridge, the density difference allows it to accelerate
2592 and entrain ambient water, practically doubling its volume. This is different from the deep

2593 convection in the Labrador Sea that increases its depth and its density only gradually and marginally
2594 relative to the underlying waters. This might be the main reason for the small contribution from the
2595 Labrador Sea to the overturning circulation.

2596 The formation of dense water in the Arctic Mediterranean also provides a constant,
2597 but perhaps weaker and often less obvious forcing than the wind driven circulation. The dense
2598 outflowing water has to be replaced, when the sea level starts to become lower, if not before. The
2599 outstanding question is: Can the processes in the Arctic Mediterranean produce enough dense, and
2600 dense enough, water in the future? If the convection does not reach to the bottom, it will create
2601 stagnant conditions in the deep basins, but as long as intermediate waters are formed, still denser
2602 than the waters in the subpolar gyre, and the pressure head at the sill is present, the overturning
2603 circulation should be maintained. A large input of freshwater to the Greenland Sea, either from the
2604 Arctic Ocean or from ice melting on Greenland, could strongly affect the convection in the Greenland
2605 Sea (Stigebrandt, 1985). Only the cooling of the Atlantic water in the Norwegian Sea and the
2606 unknown rate of deep water formation in the Arctic Ocean would then remain as sources of dense
2607 overflow water, but that might be sufficient, as it has been assumed to be before.

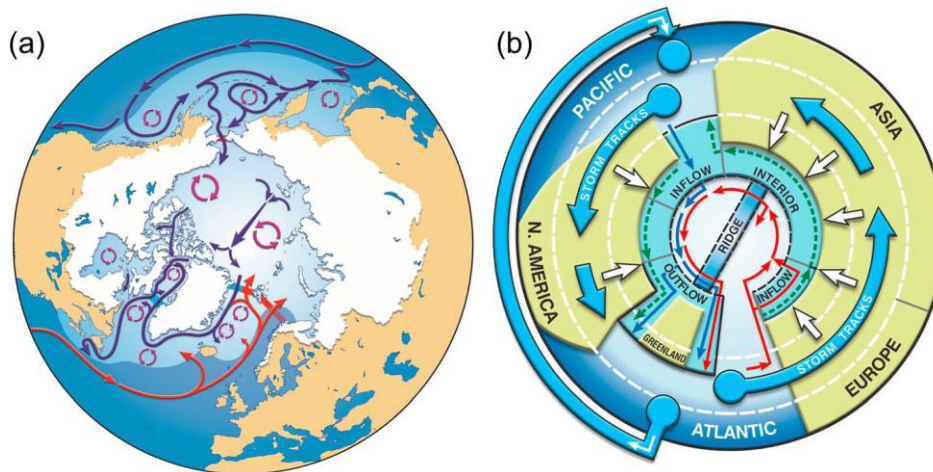
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14 Overview and Global connections and interaction.

At the global scale the Arctic Ocean is a fundamental node in the global hydrological cycle and the ocean’s thermohaline circulation (Figure 24). Here we have assessed the system’s circulation, water mass transformations, and key functions and processes. Our use of the term *Arctic Mediterranean* acknowledges that the Arctic Ocean and Nordic Seas are largely surrounded by north-flowing drainage basins supplied by moisture originally derived from subarctic seas and delivered by the Westerly wind field (cf. Carmack et al., 2008; Carmack, et al., 2016). As such they are termed *beta oceans*, where stratification is controlled by salinity which, in turn, allows ice formation and attendant seasonal fluxes of salt and freshwater (Carmack, 2007). The Arctic Mediterranean is, itself, dynamically and thermodynamically, a *double estuary*, in that its inflowing subarctic waters are transformed within into outflow waters that are both lighter and denser than their parent water masses (Carmack and Wassmann, 2006; Rudels and Carmack, 2022; Brown et al., 2025), all flowing along the east and west coasts of Greenland. There they mingle and mix with new incoming subarctic waters to assemble the lower limb of the Atlantic Meridional Overturning Circulation. The Arctic climate system, is now primed to cross multiple cascading and irreversible tipping points, as was forecast by Aagaard and Carmack (1989) who speculated, “that the present-day Greenland and Iceland seas, and probably also the Labrador Sea, are rather delicately poised with respect to their ability to sustain convection.”

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2633 Figure 24. Schematic maps of (a) the major ocean currents (long arrows), the four Arctic Ocean
2634 gateways in Fram Strait, the Barents Sea Opening, Davis Strait, and Bering Strait (thick arrows with
2635 red denoting warm and saline inflow and blue denoting cooler and fresher surface flows), the gyral
2636 circulation patterns (circular arrows), the salt-stratified ocean domains are shown in light blue, and
2637 the terrestrial regions contributing freshwater to the Arctic Mediterranean shown in white, and (b) a
2638 highly stylized summary of components of the high-latitude freshwater system. Here moisture is
2639 transported from the subtropical and tropical Atlantic Ocean to the Pacific Ocean via the Trade Winds
2640 over Central America (thick blue arrow). The subarctic front separates the thermally stratified
2641 subarctic oceans (darker blue) from the salt-stratified Northern Ocean (lighter blue) in both the
2642 Atlantic and Pacific Oceans. Moisture is transported from the Pacific and Atlantic Oceans to the Arctic
2643 catchment basins by the midlatitude (Westerlies) storm tracks (thick blue arrows), which
2644 subsequently drains into the Arctic Ocean (thick white arrows) where it spreads initially within coastal
2645 domains (dashed green arrows). Warm, salty Atlantic origin waters (thin red arrows) enter the Arctic
2646 Ocean through Fram Strait (the Fram Strait Branch) and through the Barents Sea Opening (the
2647 Barents Sea Branch) and circulate within the Arctic Basins as subsurface, cyclonic, topographically
2648 steered boundary currents along the continental margin and ridge system. Internally modified
2649 Atlantic waters exit the Arctic Ocean southward through Fram Strait along eastern Greenland. Cooler
2650 and fresher Pacific origin waters (thin blue arrows) enter the Arctic Ocean through Bering Strait and
2651 exit through the Canadian Arctic Archipelago and Fram Strait along eastern Greenland. Within the
2652 Arctic Ocean a topological distinction is made between inflow, interior, and outflow shelves [cf.
2653 Carmack and Wassmann, 2006; Bluhm et al., 2015].

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