

Constraints on exchanges in the Arctic Mediterranean—do they exist and can they be of use?

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ABSTRACT

Extensive changes have been reported from the Arctic Mediterranean. The ice cover is retreating, the temperature in the Atlantic layer has been increasing, the salinity in the upper layers shows large variations and deep waters in the Greenland Sea have become warmer and more saline. These changes all appear externally forced; by the radiation balance, by the atmosphere, and by ocean advection. The question arises—are there processes inherent to the Arctic Ocean, which can constrain changes induced by external forcing? Three features are examined; the storage and export of liquid freshwater in the upper layers, the heat loss of the Atlantic water encountering sea ice and the possibility to define a salinity separating the two roles of the Arctic Mediterranean, as estuary and as concentration basin. If the freshwater outflow in the upper layer is rotationally controlled, the liquid freshwater storage and export only depend upon the freshwater input. The melting rate of sea ice is affected both by the heat transport and by the temperature of the inflowing Atlantic water. A salinity separating the estuarine and the deep-water circulation is proposed depending upon the salinity and the temperature of the Atlantic water as it encounters sea ice.

1. Introduction

The Arctic Ocean, once remote and forbidding and demanding extreme hardship of those venturing into the north, unknown and subject to speculations, now has become the topic for headlines and a common concern. During the first 100 yr of exploration of the Arctic Ocean after the drift of Fram 1893–1896 (Nansen, 1902) the progress was slow, information scarce, and the image of a changeless, ice covered and salinity stratified ocean emerged. This image of timelessness has been seriously shattered during the last 20–30 yr, when exploration began to yield to systematic observations. The Arctic Ocean appears to be no less varying than any other ocean. The perennial ice cover was found diminishing in extent as well as in volume (Rothrock *et al.*, 1999). Several changes have also been detected in the water mass distribution in the Arctic Ocean. Pulses of warm Atlantic water have been observed in the interior of the Arctic Ocean (Quadfasel *et al.*, 1991; Polyakov *et al.*, 2005), and their pathways to and within the Arctic Ocean have been charted. The extent of the influence of the low salinity Pacific water that enters through Bering Strait was in the 1990s found to contract (Carmack *et al.*, 1995), and at the same time the freshwater input from the Siberian rivers was shifted from the Amundsen Basin

to the Makarov Basin and even to the Canada Basin (Steele and Boyd, 1998). Many of these 1990s changes were attributed to the North Atlantic Oscillation (NAO), being in a highly positive state during the 1990s (Hurrell, 1995).

The fourth International Polar Year (IPY) 2007–2009, conducted 125 yr after the first IPY in 1883, has provided a wealth of new information on the Arctic Ocean, its ice cover and its water masses, their distribution, circulation and transformation. Nature contributed significantly to the success and impact of the IPY. The Arctic Ocean ice cover attained its lowest minimum extent on record and initiated discussions about the future fate of the summer sea ice. Would the Arctic Ocean be ice free in summer in the next decade, not in the later part of this century as models have been predicting? This unexpected retreat of the summer sea ice started much speculation and also betting on the ice extent in September 2008. The minimum ice extent in 2008 was slightly larger than in 2007, but the evolution in the incoming years is open to guesses, not prediction. These drastic changes in the most easily observed feature of the Arctic Ocean have made a strong public impact, and the future of the Arctic Ocean ice cover and of the Arctic environment has become a concern, not just for local inhabitants and for the scientists involved, but for the public in general.

The observed variations have shown how little we still know about the processes active in, and controlling the state of the Arctic Ocean. 2007 was perhaps an exceptional year in the high Arctic with clear skies in early summer, persisting southerly

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winds from the Pacific sector bringing warm air over the Arctic Ocean as well as forcing large volumes of sea ice through Fram Strait southward into the Nordic Seas. But perhaps every year is special. The observations made during IPY will be examined and analysed in the near future to explain what happened in 2007 and what it might imply for the future conditions of the Arctic Ocean.

One open question is how much the variability of the Arctic Ocean can be traced back to ocean advection and changes of the oceanic heat and freshwater transports. Here we shall approach this problem by examining if, by investigating the physical properties and processes encountered in the Arctic Ocean, it is possible to identify and formulate constraints that restrict the communication between the Arctic Ocean and the rest of the world ocean as well the interactions between warm water, sea ice and the atmosphere. Constraints that could give a zero order description of the influence of oceanic advection on the conditions in the Arctic Ocean. Three features will be studied: (1) the export of low salinity surface water from the Arctic Ocean to lower latitudes; (2) the interactions between the sea ice and warmer water advected from the south and (3) the Arctic Ocean-Nordic Sea system acts as a double estuary producing both dense water, contributing to the Atlantic thermohaline circulation, and

less dense surface water, exporting the excess freshwater added to the Arctic. Is it possible from air-sea-ice interactions to define a salinity (density) surface that separates the two circulation loops?

2. The freshwater storage in the Arctic Ocean

The Arctic Ocean (Fig. 1) is an almost enclosed, high latitude sea with large net freshwater input, mainly from river runoff from the surrounding continents but also the direct net precipitation on the Arctic Ocean surface is substantial (Serreze et al., 2006). In addition, freshwater is advected to the Arctic Ocean from the Pacific Ocean. The Pacific is less saline than the Atlantic and this imbalance is partly corrected by a flow of less saline upper layer water, driven by the higher sea level in the Pacific, through the 50 m deep Bering Strait into the Arctic Ocean (Stigebrandt, 1984). Finally also the Norwegian Coastal Current contributes a smaller volume of freshwater to the Arctic Ocean (Dickson et al., 2007).

The freshwater input leads to strong stability in the upper part of the Arctic Ocean water column that limits the local convection to a shallow surface layer. Heat is removed by cooling in winter and the surface layer temperature decreases to the freezing point

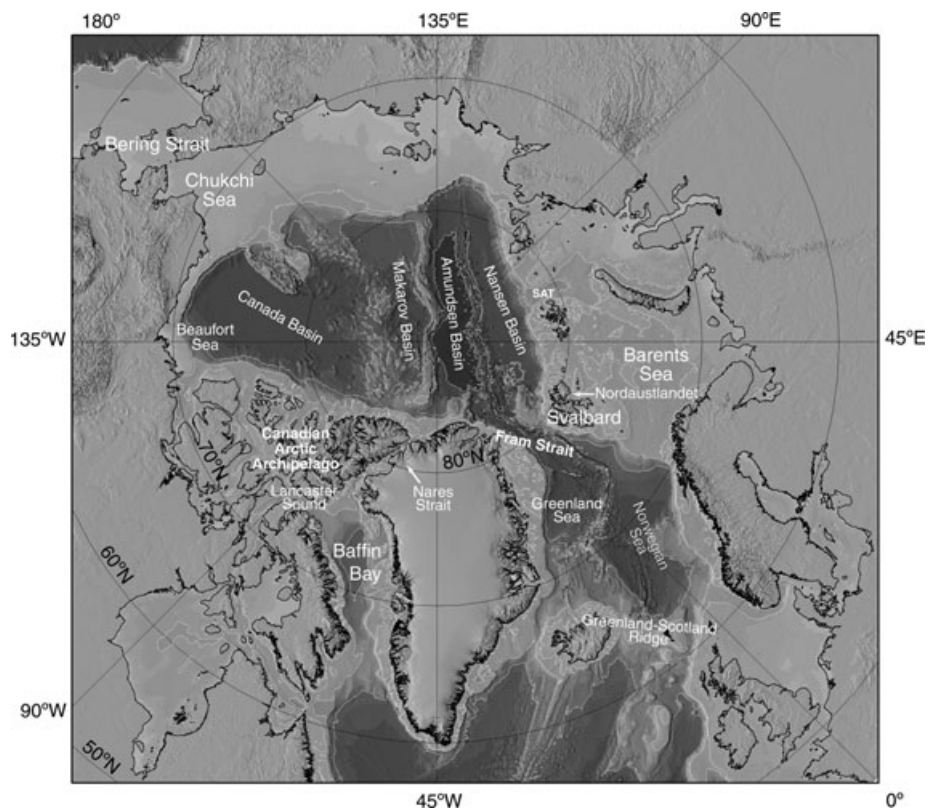


Fig. 1. Map of the Arctic Mediterranean Sea showing the outflow passages in the Canadian Arctic Archipelago and Fram Strait and the location of the main bathymetric features mentioned in the text. The bathymetry is from Jakobsson et al. (2008).

and ice is formed. The freshwater input and the strong stability is thus crucial for the formation and survival of the ice cover by inhibiting the vertical heat transfer from the underlying, warm Atlantic water to the sea surface and to the ice and atmosphere.

It is then obvious that the freshwater balance of and the freshwater storage in the Arctic Ocean are of importance for the existence of the Arctic Ocean ice cover and for the Arctic environment in general. A loss of the permanent ice cover is arguably the most significant single change that could occur in the Arctic. In the past observations have not been dense enough in space and frequent enough in time to make accurate estimates of the freshwater storage in the Arctic Ocean, not to mention its variability. Most of the information of the variability of the freshwater storage during the last 50 yr has been derived from model results rather than from direct observations (e.g. Häkkinen and Proshutinsky, 2004). With the increase in observations in recent years provided by ice breaker expeditions and by the deployment of ice-tethered platforms and from the intense observational efforts made during IPY it will most likely be possible to obtain reliable estimates of the freshwater storage and its variations from direct observations in the near future.

We are not there yet, and we shall explore another, much simpler, complementary approach and try to determine the liquid freshwater storage in the Arctic Ocean required to establish a balance between the freshwater input and the freshwater export. We assume that the liquid freshwater transport takes place in an upper low salinity layer lying above more saline and denser water, which to a zero order approximation is taken to be at rest. The flow in the upper layer is assumed to be geostrophic and the transport controlled by the conditions at the outflow passages, Fram Strait and the narrower channels in the Canadian Arctic Archipelago (Fig. 1).

We use the expression, eq. (1), derived by Werenskiöld (1935) to determine the transport (see also Defant, 1961). Werenskiöld considered a low salinity wedge flowing along the coast above a lower layer at rest. He found that if the lower layer reached the surface somewhere away from the coast the transport M in the wedge could be estimated from the density difference between the layers and the depth H of the upper layer at the coast. For a homogenous upper layer with density ρ_1 the expression becomes:

$$M = \frac{(\rho_2 - \rho_1) g H^2}{2\rho_2 f}, \quad (1)$$

Here f is the Coriolis frequency and g the acceleration of gravity and ρ_2 the density of the lower layer.

The outflow takes place with the coast to the right and the lower layer should reach the surface within the strait to have full transport capacity. For the six passages in the Canadian Arctic Archipelago we postulate that they can be approximated by two straits transporting at full capacity. A similar approach was used by Stigebrandt (1981a). Together with Fram Strait

this means that the liquid freshwater is exported by a baroclinic geostrophic flow through three passages. The ice export, by contrast, is considered decoupled from the liquid freshwater export, being mainly driven by the wind. We shall return to the effects of wind and to the importance of barotropic exchanges after the implications of the baroclinic transport of liquid freshwater have been explored.

We compute the freshwater content in the upper layer relative to the salinity of the lower layer S_A , which is assumed the same in all channels and set to 35, the salinity of the Atlantic water. This implies that the Pacific water is regarded as comprised by freshwater and Atlantic water entrained into the upper layer. The liquid freshwater balance requires that the exported freshwater volume F is equal to $F_o - F_i$, where F_o is the total freshwater input to the Arctic Ocean and F_i the freshwater exported as ice. The volume transport through the three channels is then given by

$$M_A + F = \frac{3g\beta(S_A - S_1)H_1^2}{2f}. \quad (2)$$

Here S_1 and H_1 are the salinity and depth of the upper layer and M_A the amount of Atlantic (lower layer) water entrained into the surface layer, $g = 9.83 \text{ m}^1 \text{ s}^{-2}$ the acceleration of gravity and $f = 1.4 \times 10^{-4} \text{ s}^{-1}$ the Coriolis frequency. To further simplify the approach and illuminate the results we have followed Stigebrandt (1981a) and adopted a simplified equation of state taking into account only the effects of salinity on the density.

$$\rho = \rho_f(1 + \beta S). \quad (3)$$

Here ρ_f is the density of freshwater and β the coefficient of salt contraction set equal to 8×10^{-4} . With $\rho_2 = \rho_f = 1000 \text{ kg m}^{-3}$ the reduced gravity can be written as

$$g' = g \frac{\rho_2 - \rho_1}{\rho_2} = g\beta(S_A - S_1). \quad (4)$$

The freshwater content m (in metre of freshwater) in the upper layer is

$$m = \frac{(S_A - S_1)H_1}{S_A}. \quad (5)$$

The salinity S_1 in the upper layer can be written as

$$S_1 = \frac{M_A S_A}{(M_A + F)} \quad (6)$$

and if S_1 is introduced in the expressions for the outflow and for the freshwater content we obtain

$$(M_A + F)^2 = \frac{3g\beta F S_A H_1^2}{2f} \quad (7)$$

and

$$m = \frac{F H_1}{(M_A + F)}. \quad (8)$$

If we assume that the maximum upper layer depth in the channels is equal to the depth of the upper layer in the interior of the Arctic Ocean this gives the freshwater storage in the Arctic Ocean as

$$m = \left(\frac{2fF}{3g\beta S_A} \right)^{1/2}. \quad (9)$$

Introducing the fairly realistic values $F_o = 0.29 \times 10^6 \text{ m}^3 \text{ s}^{-1}$, $F_i = 0.09 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ leading to $F = 0.2 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ (Dickson et al., 2007), we find a freshwater storage of 8.25 m. If the ice in the Arctic Ocean were only winter ice, melting in summer, and no ice were exported, the freshwater input F would increase by 0.09 Sv ($1 \text{ Sv} = 1 \times 10^6 \text{ m}^3 \text{ s}^{-1}$) and the freshwater storage would become 10.1 m. The freshwater storage in the upper layer thus increases as $F^{1/2}$, due to an increased freshwater input. However, the residence time of the freshwater in the Arctic becomes shorter as $F^{-1/2}$.

The Rossby radius (R_o) is found to be independent of the depth of the upper layer and of the entrainment and the total transport and only affected by the freshwater storage and the salinity of the lower layer.

$$R_o = \frac{[g\beta(S_A - S_1)H_1]^{1/2}}{f} = \frac{(g\beta S_A m)^{1/2}}{f} = \left(\frac{2g\beta S_A F}{3f^3} \right)^{1/4}. \quad (10)$$

For the present situation with three outflow channels this corresponds to a Rossby radius of $\sim 10\,800$ m. This is narrow enough to allow the upper layer water to pass through both Lancaster Sound and Nares Strait in the Canadian Arctic Archipelago without the upper layer extending across the channels.

These results show that the freshwater export and the Rossby radius are determined solely by the thickness of the freshwater layer and not influenced by the entrainment of Atlantic water. However, the average thickness of the ice cover, 2–3 m, would, if the ice were floating on an only 8-m-thick freshwater layer, significantly increase the thickness of the upper layer by displacing freshwater and thus contribute to the freshwater export. In fact, the ratio of the ice thickness to the liquid freshwater content, that is, 2/8 could be used as a first estimate of the Arctic Ocean sea ice exported by the baroclinic ocean currents. This would amount $1/4$ of the liquid freshwater export, rather than close to $1/2$ as has been assumed here based on Dickson et al. (2007). Since the upper layer is about ten times thicker than the pure freshwater layer the dynamic effect of the ice cover can be ignored and only the liquid freshwater content needs to be considered.

The present approach neglects all effects of barotropic exchanges on the freshwater export. How serious is this neglect? The barotropic transport can be separated into two categories, barotropic, bathymetrically steered exchanges in the deep Fram Strait, and transports due to piling up of the sea surface caused by local winds. The latter involves the accumulation of low

salinity surface water towards the Greenland coast in Fram Strait and the generation of sea level differences between the Beaufort Sea and Baffin Bay across the Canadian Arctic Archipelago (Prinsen et al., 2009). For the deep barotropic exchanges through Fram Strait we note that most freshwater is carried by the upper water masses (see e.g. Häkkinen and Proshutinsky, 2004; Rudels et al., 2008) and the only possibility for the barotropic velocities to increase the freshwater export is by adding a barotropic velocity component to the baroclinic velocity in the upper layers. Although the barotropic velocities, because of the large depth of Fram Strait, contribute significantly to the total volume exchanges, their addition to the freshwater transport would be comparatively small because the baroclinic velocities in the upper layers are clearly larger than, for example, a barotropic velocity capable of doubling the total exchanges through Fram Strait.

When the barotropic flow is generated by a wind induced sea surface slope, either at the Greenland coast or between the Beaufort Sea and Baffin Bay, we expect this transport to be added to the baroclinic transport. If the sea level slope induces an increased freshwater export, this will drain the freshwater pool in the Arctic Ocean, and should this situation prevail over a longer time the upper layer depth and the baroclinic contribution will gradually diminish until the combined outflow balances the freshwater input. Since the barotropic transports induced locally by wind as well as added by bottom steered flows forced by larger scale meteorological conditions are expected to vary on time scales shorter than that of the freshwater supply, their effects can be regarded as perturbations on a basic baroclinic transport.

The ice export has been assumed decoupled from the baroclinic transport and driven by the wind, either directly through the Ekman drift or by the piling of water against the Greenland coast. Observations have shown that almost all (>90%) of the ice export occurs through Fram Strait (Serreze et al., 2006; Carmack et al., 2008). It was estimated above that the ice export carried by the baroclinic flow through the three straits would amount to $1/4$ of the liquid freshwater export. One-third of the liquid freshwater export takes place in Fram Strait and since the total ice export (0.09 Sv) is slightly less than half the total liquid freshwater outflow (0.2 Sv) the liquid freshwater export through Fram Strait is about two third of the ice export through Fram Strait. This supports the assumption that the ice export is largely independent of the baroclinic freshwater outflow and primarily driven by the wind.

Another assumption is that the lower layer has the same salinity ($S = 35$) in all passages. This salinity is appropriate for Fram Strait but is too large for the Canadian Arctic Archipelago, where the salinity difference is between the Beaufort Sea and the upper layer of Baffin Bay, which has a salinity of ~ 33.7 (e.g. Rudels, 1986). The estimated baroclinic transport through the Canadian Arctic Archipelago is then likely too large compared to that in Fram Strait because of the smaller density difference. Another factor to consider is the depth of the channels in the Canadian

Arctic Archipelago. Even if the Rossby radius is independent of the depth of and transport in the upper layer, the sill depth might interfere and reduce the transport, if the lower part of the upper layer cannot pass over the sill. The depth of the upper layer has been taken to be the same in all channels and equal to the depth of the upper layer in the Arctic Ocean. This is not in agreement with observations. The upper layer in the Eurasian Basin is thinner and more saline than the upper layer in the Canadian Basin that contributes to the flow through the Canadian Arctic Archipelago. Since these variations act opposite to the density difference between the two layers, they may to some extent compensate the assumption of equal characteristics of the lower layer in all channels. These reflections just illuminate the zero order nature of this approach.

The large scale meteorological forcing, driven, for example, by the different states of the Arctic Oscillation (AO) and the North Atlantic Oscillation (NAO), will also affect the depth of the upper layer differently in the Canadian Arctic Archipelago and in Fram Strait. A negative AO state would strengthen the anticyclonic circulation in the Beaufort Gyre and could lead to a deeper upper layer in the Canadian Arctic Archipelago than in Fram Strait. Since the transports vary as the square of the upper layer depth this implies a larger total liquid freshwater export to the North Atlantic, because the increase of the outflow through the Canadian Arctic Archipelago would be larger than the reduction of the export through Fram Strait. The opposite situation would occur when the AO index is positive. The freshwater and the upper layer depth would be more equally distributed between the different Arctic Ocean basins and the upper layer thickness in the passages would be more similar. Since the minimum export occurs when the thicknesses of the upper layer in all passages are equal, the freshwater export would increase in Fram Strait but the total freshwater export to the North Atlantic would decrease. This, however, requires that the freshwater during the AO—state is not accumulated in an isolated pool in the Beaufort Sea as presently appears to take place (Andrey Proshutinsky, personal communication). In that case the upper layer thickness also decreases in the channels of the Canadian Arctic Archipelago (or increases less) and the total export becomes smaller. Again, once the gyre relaxes, the depth of the upper layers in the channels increases and the freshwater export rises momentarily above the equilibrium baroclinic transport, until the freshwater storage again attains its balance between freshwater input and freshwater export. An excessive rise in sea level in the Beaufort Sea and the southern Canada Basin could also influence the inflow of Pacific water by reducing the driving sea surface slope in Bering Strait.

3. The oceanic sensible heat flux to the Arctic Ocean

The meridional northward transport of heat carried by the ocean is estimated to be a considerable fraction of the atmospheric

heat transport also at high latitudes (see e.g. Peixoto and Oort, 1992, figure 13.17). This in spite of the narrow passages that connect the Arctic Ocean with the world ocean, Bering Strait in the North Pacific and the Nordic Seas in the North Atlantic. Of these two passages the Nordic Seas, carrying about 10 times the transport through Bering Strait, has commonly been considered the most important one for the Arctic Ocean heat balance. In particular the inflow of warm Atlantic water through Fram Strait, the only deep passage to the Arctic Ocean, has been regarded as critical. However, northward transport of warm water is not enough to provide heat to the Arctic. The heat has to be given up to the environment and eventually contribute to the long-wave back radiation to space required to balance the global heat budget.

3.1. Melting of sea ice north of Fram Strait

The strong stability of the Arctic Ocean caused by the large freshwater input inhibits the heat residing in the deeper layers from reaching the sea surface, the ice and the atmosphere. The question then arises—what is the efficiency of the oceanic heat transport to the Arctic Ocean? The crucial areas are the Nansen Basin immediately northeast of Fram Strait and the northern Barents Sea, where the warm Atlantic water, which is at the surface in the Norwegian Sea, becomes transformed into a subsurface warm core, supplying the 300–600 m thick Atlantic layer with temperatures above 0 °C in the Arctic Ocean, and a cold, less saline surface layer. In the Barents Sea the Atlantic water is cooled significantly before it reaches the Arctic Ocean, sinking down the St. Anna Trough. The main inflow of warm Atlantic water thus takes place through Fram Strait. Here the Atlantic water encounters and melts sea ice. In summer this melting is augmented by heat input from above, either by short-wave solar radiation directly on the ice or by heating the low salinity melt water dominated surface layer, leading to basal melting of the ice. In winter the only heat source is the Atlantic water. The melting is strong enough to keep the area north of Svalbard almost permanently free of ice as far east as Nordaustlandet. This region, the Whalers' Bay, is where the main direct interactions between warm Atlantic water and sea ice take place.

The melting of sea ice and the mechanical stirring supplied by the wind create a surface layer comprising sea ice melt water and Atlantic water, which is cooled towards freezing temperature. If all heat in a layer of thickness H_1 is given up to ice melt the amount of melt water m_m added to the water column is

$$m_m = \frac{cH_1\Delta T_A}{L}, \quad (11)$$

where $c = 4000 \text{ Jkg}^{-1} \text{ K}^{-1}$ is the specific heat of seawater and $L = 0.335 \times 10^6 \text{ Jkg}^{-1}$ the latent heat of fusion and ΔT_A the reduction of temperature in the cooled Atlantic water. The

salinity of the upper layer then becomes

$$S_1 = \frac{S_A H_1}{H_1 + m_m} = \frac{S_A}{(1 + c\Delta T_A L^{-1})}. \quad (12)$$

For a temperature reduction of 4° , from $+2^\circ\text{C}$ to the freezing point and $S_A = 35$, this gives a salinity of ~ 33.4 .

Observations indicate that the salinity of the upper layer is higher than it would be if all heat given up by the Atlantic water is used to melt ice, and a large fraction, 60–75%, must be lost to the atmosphere to account for the high salinity (Fig. 2). We accept that such heat loss to the atmosphere occurs without trying to specify how the actual transfer takes place, in leads or by conduction through the ice. The heat loss then consists of two parts, a fraction ϕ going to ice melt and the remainder $(1 - \phi)$ going to the atmosphere, and we shall try to determine ϕ_o as the fraction giving minimum ice melt.

The region of interactions is considered large enough for the horizontal distribution of ice and open water to be statistically homogenous and that the vertical heat transfer from the Atlantic water to the ice can be described by a one-dimensional Kato–Phillips energy balance model, where the turbulent energy supplied by the wind is parametrized by a friction velocity u_* (Kato and Phillips, 1969; Stigebrandt, 1981a). Because of ice melt buoyancy is added to the sea surface and a part of the energy must be used to mix the melt water into the upper layer. In this process the temperature contribution to the density is crucial and we cannot work with the simplified equation of state used above (eqs 3 and 4). Following Rudels et al. (1999) we have the

entrainment velocity w_e from below given by

$$w_e = \frac{2n_o u_*^3}{g[\beta(S_A - S_1) - \alpha(T_A - T_1)]H_1} - \frac{\varepsilon B}{g[\beta(S_A - S_1) - \alpha(T_A - T_1)]}, \quad (13)$$

where T_A and T_1 are the temperatures of the Atlantic and upper layer respectively, H_1 is again the thickness of the upper layer, n_o is a constant = 1.25 and α is the coefficient of heat expansion taken to be a function of temperature. B is the buoyancy flux and ε is a constant set equal 1 when the buoyancy input is positive and 0.05 when the buoyancy flux is negative (Stigebrandt, 1981a).

We assume that as the Atlantic water encounters sea ice the upper layer immediately reaches a depth large enough to have a balance between the heat loss at the upper boundary and the entrainment of warm water through the interface between the upper layer and the Atlantic water below. This leads to a small temperature and salinity reduction of the upper layer, comprising Atlantic water and melt water. As the cooling proceeds and the mixed layer deepens entrainment from below cannot balance the heat loss at the surface and the temperature of the upper layer decreases. As the heat is lost both to ice melt and to the atmosphere the salinity of the upper layer is also reduced.

The buoyancy input B has two terms, one negative due to the heat loss Q of the Atlantic water in the upper layer as well as the Atlantic water entrained from below, and one positive caused by the freshwater added by the melting of sea ice. The melting uses the fraction ϕQ of the sensible heat loss Q of the Atlantic water.

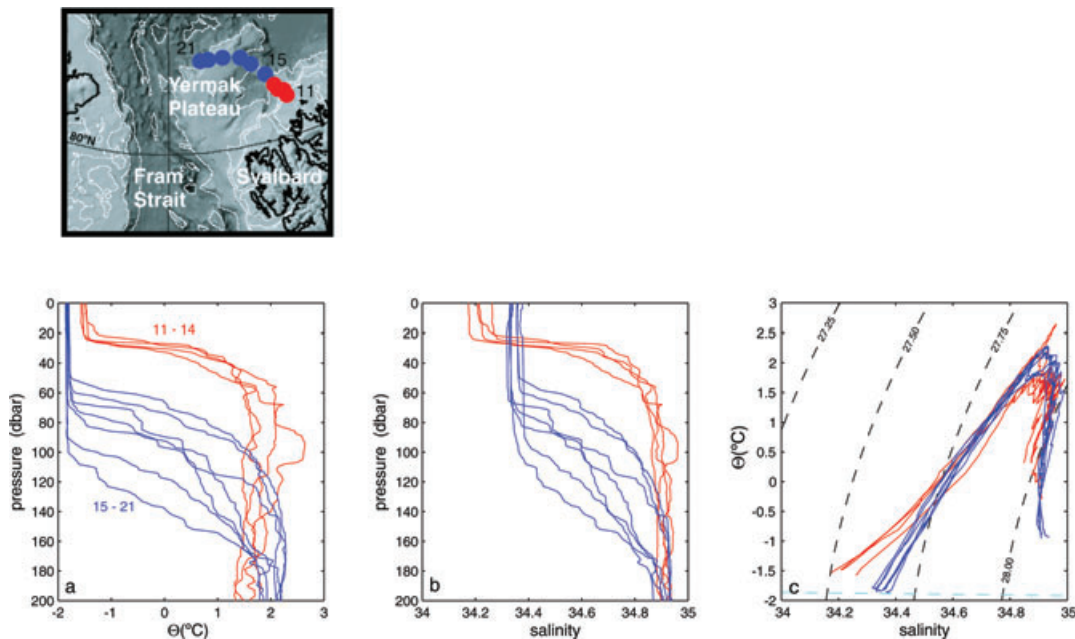


Fig. 2. The Atlantic water that enters the Arctic Ocean through Fram Strait encounters and melts sea ice north of Svalbard. The salinity of the upper layer is too high, if all the sensible heat in the cooled Atlantic layer is going to ice melt. Some heat must also be lost to the atmosphere. The stations are from the Oden AO-02 expedition and the figure is adapted from Rudels et al. (2005).

Relating Q to the entrainment and cooling of the Atlantic water, the buoyancy flux into the upper layer becomes (for this and the following see Rudels et al., 1999 for details)

$$B = g \left\{ \left[w_e \frac{\phi c}{L} \Delta T_A \beta S_A + \frac{\phi c}{L} \beta S_A H_1 \frac{d(\Delta T_A)}{dt} \right] - \left[w_e \alpha \Delta T_A + H_1 \alpha \frac{d(\Delta T_A)}{dt} \right] \right\}. \quad (14)$$

Here again ΔT_A is the temperature reduction of the Atlantic water in the upper layer. By introducing the expression for the buoyancy flux into the equation for the entrainment velocity with $\varepsilon = 1$ (melting conditions) we obtain

$$w_e = \frac{n_o u_*^3 (1 + \phi c L^{-1} \Delta T_A)}{g \Delta T_A [\phi c L^{-1} \beta S_A - \alpha (1 + \phi c L^{-1} \Delta T_A)] H_1 - \frac{H_1}{2 \Delta T_A} \frac{d(\Delta T_A)}{dt}}. \quad (15)$$

The sensible heat lost by the Atlantic water can then, after considerable algebra, be written as

$$Q = G \left\{ u_*^3, \Delta T_A, t, \frac{d(\Delta T_A)}{dt} \right\} \times \frac{1}{\sqrt{\phi c L^{-1} \beta S_A - \alpha (1 + \phi c L^{-1} \Delta T_A)}}. \quad (16)$$

Here $G\{\}$ is a functional expression of the bracketed terms that does not depend upon ϕ .

In case that the reader should turn to Rudels et al. (1999) for clarification, it should be noted that the corresponding eq. (20) contains a mistake in the expression for G . Eq. (20) should read, in the notation used there

$$Q = c \rho \sqrt{\frac{m_o u_*^3}{g}} \times \left(\sqrt{\frac{\Delta_D T}{2t}} + \sqrt{\frac{t}{2 \Delta_D T} \frac{d(\Delta_D T)}{dt}} \right) \times \frac{1}{\sqrt{f c L^{-1} \beta S_D - \alpha (1 + f c L^{-1} \Delta_D T)}}.$$

(This was pointed out to me by Johanna Nilsson, University of Göteborg. Since only the last factor is used in the subsequent analysis, this mistake does not affect the results.)

Eventually the upper layer reaches freezing temperature and ice formation will be necessary to fulfil the heat transfer through the upper boundary. We postpone the discussion of this stage until Section 4 and concentrate on the situation up to the time, when the surface layer reaches the freezing point. The aim is to determine the fraction ϕ going to ice melt, and to do this we assume the existence of a cooperative atmosphere that absorbs all heat it receives. The control of the vertical heat transport is then given by the ice melt, which determines the stability at the lower boundary and thus the entrainment velocity. We search for the fraction of heat going to ice melt that leads to the lowest melting rate, and thus creates the least obstruction to the vertical

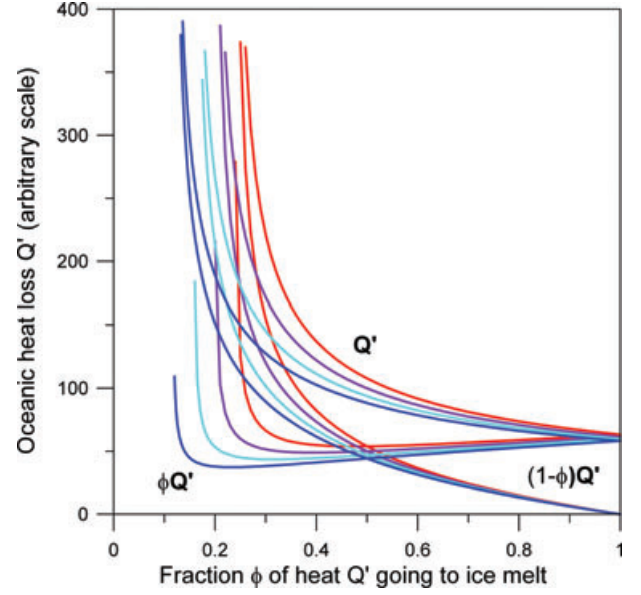


Fig. 3. Curves showing the total heat loss Q' (from the second factor in eq. 16), the heat going to ice melt $\phi Q'$, and the heat given up to the atmosphere $(1 - \phi) Q'$ as functions of ϕ , the fraction of heat going to ice melt. Red curves have $\Delta T_A = 8^\circ \text{C}$ and $\alpha = 0.781 \times 10^{-4}$, purple curves $\Delta T_A = 6^\circ \text{C}$ and $\alpha = 0.654 \times 10^{-4}$, cyan curves $\Delta T_A = 4^\circ \text{C}$ and $\alpha = 0.526 \times 10^{-4}$ and blue curves $\Delta T_A = 2^\circ \text{C}$ 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 the minimum ice melt decreases with decreasing temperature and decreasing α .

heat transfer. That ϕ_0 is a minimum is seen in Fig. 3 which shows the second factor, here denoted Q' , in eq. (16) as well as $\phi Q'$ and $(1 - \phi) Q'$. To find ϕ_0 we form

$$\frac{d(\phi Q')}{d\phi} = 0, \quad (17)$$

where Q' again denotes the factor from eq. (16) that depends upon ϕ . We then find:

$$\phi_0 = \frac{2\alpha L}{c(\beta S_A - \alpha \Delta T_A)} \approx \frac{2\alpha L}{c\beta S_A}. \quad (18)$$

For the situation when $\phi > \phi_0$ the situation is easy to visualize. If more heat goes to melting, the stability at the lower boundary increases, less heat is brought into the upper layer and the melting rate is reduced. The stability then decreases and the heat transfer from below increases and the system moves back towards the state with minimum melt rate.

If the fraction going to ice melt is less than ϕ_0 , the melt rate increases towards smaller ϕ . This implies that when the fraction going to ice melt decreases, the heat loss to the atmosphere lowers the stability sufficiently to increase the entrainment rate and bring more heat into the upper layer to further increase the melt rate. It is seen that when $\phi = 1/2 \phi_0$ the denominator of the first term in eq. (15) goes to zero. The melting does not supply

enough buoyancy to the entrained water to compensate for the buoyancy loss due to cooling and the water column overturns. In such situation, the ice cover has hardly any effect on the heat loss, which is going almost completely to the atmosphere. This might be the case in the outer part of the marginal ice zone, where the ice is scattered in mainly open water and rapid melting of the ice can take place at the same time as most of the heat is given up to the atmosphere.

Here we concentrate on the situation, where the ice cover is compact enough to obstruct the heat transfer to the atmosphere and that the ice melt is large enough to reduce the heat flux to the mixed layer from below. We postulate that in this situation the fraction going to ice melt is given by ϕ_o and once ϕ_o is determined, it can be used to estimate the heat flux from the Atlantic water going to ice melt and to the atmosphere, when the vertical heat flux Q_o at the upper boundary is forced by and determined from observed wind velocities and air temperatures and from the radiation balance. According to eq. (18) ϕ_o depends upon α , the coefficient of heat expansion, which is a function of temperature and highly non-linear at low temperatures, ranging from 0.254×10^{-4} at -2°C to 1.021×10^{-4} at 4°C . When the inflowing Atlantic water is warmer, this implies not only that more heat is released as it is cooled to freezing temperature but also that more of the lost heat is going to ice melt. The effect of a higher temperature of the northward flowing water in the West Spitsbergen Current would then be to increase the ice melt and the area of open water north of Svalbard more than can be anticipated from just the higher temperature in the Atlantic water. The changes are larger the closer to the freezing temperature the Atlantic water temperature initially is. However, southward ice

drift, caused by strong northerly winds, could easily mask this effect on the ice distribution.

The salinity of the upper layer as it reaches the freezing temperature is also determined by ϕ_o and thus from the temperature and salinity of the Atlantic layer. A higher temperature of the Atlantic layer leads to lower salinity in the upper layer and to stronger stability. This will be explored in detail in Section 4. Further cooling when the upper layer is at freezing temperature leads to ice formation and the salinity begins to increase above this ‘equilibrium’ salinity because of brine rejection. However, if the ice cover is fairly compact we do not expect too large deviations from the equilibrium salinity in the upper layer during a winter season.

3.2. Heat exchange in the interior of the Arctic Ocean

In the interior Arctic Ocean summer melting caused by the surface heating creates a low salinity, slightly warmer surface layer above a more saline temperature minimum that is the remnant of the convection the preceding winter (Fig. 4). The existence of this minimum indicates that heat is not transferred from the layers below the minimum to the surface water in summer and the heat causing ice melt is due to the incoming solar radiation. In winter the melt water layer is cooled to freezing, ice is formed and the rejected brine increases the salinity and the upper layer is eventually homogenized down to the temperature minimum and cooled to freezing temperature, creating the polar mixed layer (PML). Heat supplied from below cannot provide all heat lost at the upper boundary, and ice is formed and the salinity of the upper layer increases above the “equilibrium” value

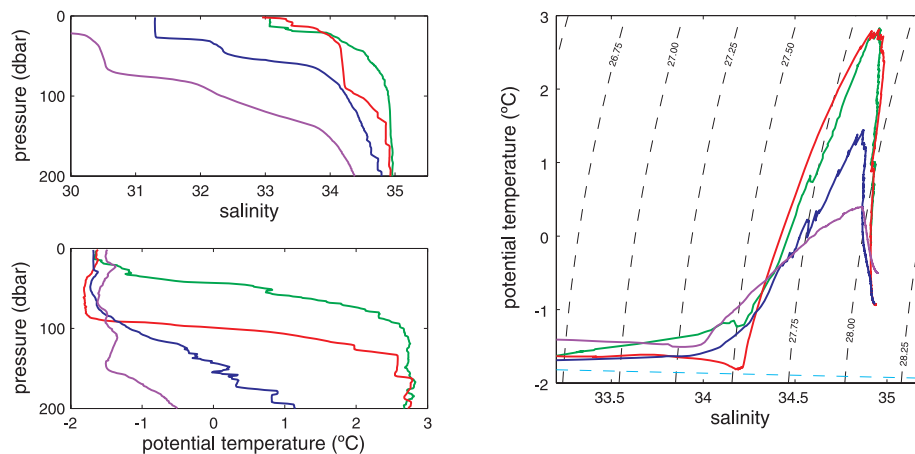


Fig. 4. In the interior of the Arctic Ocean the upper layer is commonly separated from the warm Atlantic water by a cold halocline (the blue station from the Amundsen Basin and the magenta station from the Makarov Basin). Only in the Nansen Basin (red station) do the thermocline and halocline coincide, but the stirring in the upper layer during winter is primarily due to freezing, brine rejection and haline convection. The entrainment of heat, should it become large, then removes its own energy supply. At the continental slope west of Franz Josef Land (green station) the Atlantic water comes close to the sea surface and the temperature minimum, a remnant from the winter homogenization seen on the other stations, is not always present. This indicates that entrainment takes place also in summer, when the stratification is strong, and in winter when the stratification is weaker this could reduce the ice formation.

resulting from a melt rate of $\phi_o Q$ (see also eq. 19 below). The heat loss from the Atlantic layer is expected to be small in the interior of the Arctic Ocean, where a strong, cold halocline is present between the PML and the thermocline reaching to the temperature maximum of the Atlantic layer. Only in the Nansen Basin do the halocline and the thermocline coincide and a direct contact between the winter mixed upper layer and the Atlantic water exists, but also here the entrainment of heat from below is expected to be small because the large depth of the upper layer (>100 m) (Fig. 4). The situation is different over the continental slope and the shelf break. Here the Atlantic water approaches the sea surface and observations indicate that the temperature minimum from the homogenization the preceding winter can be absent (Fig. 4). This implies that heat is entrained from the Atlantic water into the upper layer and reaches the ice and the atmosphere also in summer, and it could contribute significantly to the heat balance of the Arctic as the upper layer is cooled and homogenized during the subsequent winter.

Since the Atlantic water lies deep and is covered by a cold stabilizing halocline in most of the Arctic Ocean, its contribution to ice melt in the interior of the Arctic Ocean is small. In this respect the Pacific water may be more important. The strongest Pacific inflow occurs in summer, when its temperature is high (Coachman and Aagaard, 1988), the Bering Sea Summer Water (BSSW; Coachman and Barnes, 1961). In recent years this heat has not been removed by cooling in fall but has remained as a temperature maximum residing between 50 and 100 m depth, much shallower than the Atlantic water. It has been proposed that this heat can be brought to the sea surface by upwelling effects caused by the location of the ice edge beyond the continental slope and by interactions with the general circulation of the Beaufort gyre and thus contribute to the ice melt (Shimada et al., 2006).

Here we suggest a one-dimensional process, which also might increase the vertical heat flux from the BSSW to the ice. If the heat residing in the BSSW is removed in fall, the ice formation will be delayed because the heat initially lost by the ocean is sensible heat. When freezing finally starts the air temperatures are, in general, lower than in early autumn and a rapid ice formation will occur because of the initially open water. If a strong stability in the water column allows ice to form in autumn before the heat is removed, the freezing period becomes longer but the temperatures are initially higher, and when the winter sets in the ice cover is thick enough to reduce the heat loss and the ice formation rate. By the end of winter there will not be much difference between the total amounts of ice formed, whether the heat has been removed or not. However, if the heat is not removed in fall, the stability above the warmer layer will be lower by the end of winter because of the brine rejection, and the heat will now be more easily entrained into the surface layer. The entrainment will take place at the same time as the sun returns and the polar day begins. The ice is heated and starts to melt from both sides. This timing of the heat input from below could

cause an accelerating ice melt and contribute to a reduction of the ice cover. The minimum sea ice extent has been reduced significantly in recent years, especially in the Canada Basin north of the Chukchi Sea, where the BSSW is usually present and where a strong temperature maximum has been observed at 50–100 m depth. The Pacific water rather than the Atlantic water would then provide the oceanic sensible heat flux that mostly affects the ice cover and the conditions in the Arctic.

4. A double estuary

The Arctic Ocean and the Nordic Seas—the Arctic Mediterranean, in addition to being areas of large freshwater input, are areas subject to strong cooling, leading to dense water formation and convection, ventilating the deep basins of the Arctic Mediterranean and contributing significantly to the thermohaline overturning of the Atlantic. The Arctic Mediterranean is a double estuary, acting both as a concentration (Mediterranean) basin and as a dilution (fjord) basin. Fig. 5 present different stations showing the effects of the main processes, the dilution of the surface water, the brine rejection on the shelves, the entrainment into the water sinking down the slope and the deep open ocean overturning in the Greenland Sea, on the water column. Available observations indicate that at present the deep overturning is the dominant circulation loop and about $3/4$ of the Atlantic water passing north of the Greenland–Scotland Ridge returns as colder, slightly less saline and denser overflow water (Hansen and Østerhus, 2000). Our concern here is not to explore the stability of the present circulation but to identify a salinity that could be used to separate the less dense surface loop from the denser deep-water circulation and to find on what factors such salinity could depend.

We explore the interactions occurring when the warm, advected water encounters sea ice under freezing conditions. The freshwater input is restricted by the amount of sensible heat carried by the water and available for ice melt. If we use ϕ_o from eq. (18) as the fraction of heat going to ice melt the salinity of the upper layer will, as it reaches freezing temperature, have the salinity S_1 given by

$$S_1 = \frac{S_A}{\left(1 + \frac{\phi_o c \Delta T_A}{L}\right)} \approx \frac{S_A}{\left(1 + \frac{2\alpha \Delta T_A}{\beta S_A}\right)}. \quad (19)$$

We propose that S_1 is used as the salinity and the freezing point as the temperature defining the density surface separating the less dense surface loop from the overflow circulation. The salinity depends both on the salinity and on the temperature of the advected water as well as on the coefficient of heat expansion α . A lower temperature leads to less ice melt and this is accentuated by the smaller α at lower temperatures.

To examine if such a choice agrees with observations the salinity of the upper mixed layer observed in the Fram Strait and the Barents Sea inflow branches to the Arctic Ocean are

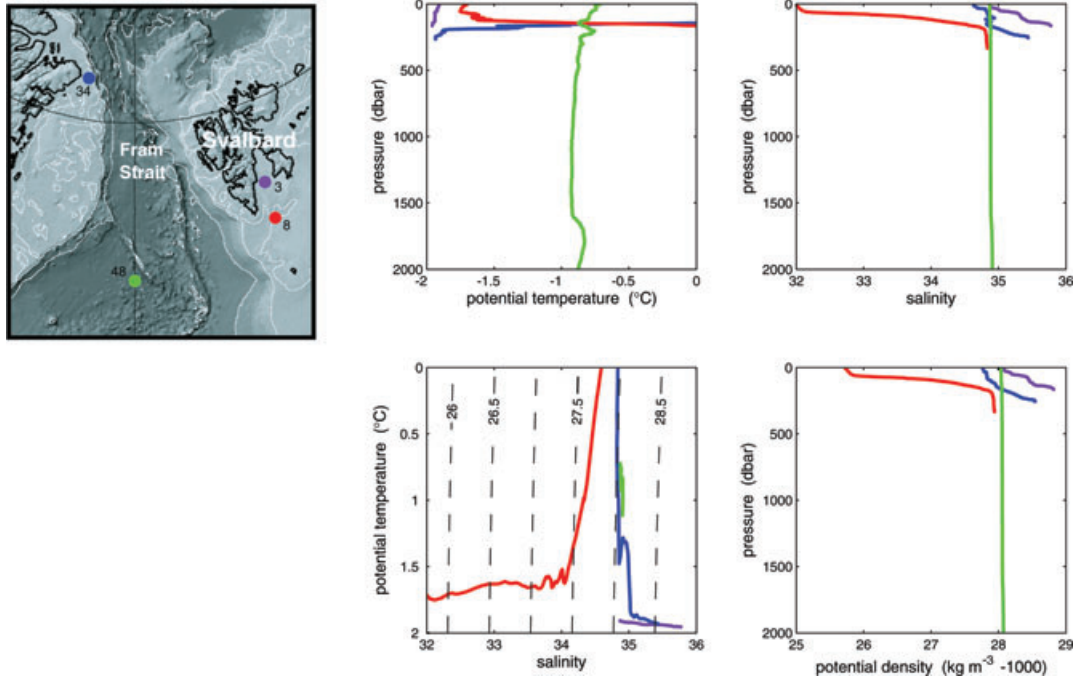


Fig. 5. Different processes showing the characteristics of the Arctic Mediterranean as a double estuary: The blue station from the East Greenland Current shows the freshening of the upper layers in the Arctic Ocean. The violet station indicates brine rejection and brine accumulation in Storfjorden. The red station shows slope convection in Storfjordrenna. The green station is an example of open ocean deep convection in the Greenland Sea. The stations are from the Oden AO-02 expedition, and the figure is adapted from Rudels et al. (2005).

compared. Both inflow branches encounter sea ice before they are significantly influenced by river runoff, and ice melt can be considered the only freshwater source. A similar situation is present in the Weddell Sea in the Southern Ocean and the salinity of the winter mixed layer should also there be controlled by the sea ice melt. All three cases indicate that the salinity is

higher than that expected, if all heat lost were used to melt ice, eq. (12). The salinities at the base of the upper layer, corresponding to the winter conditions, also follow the expression, eq. (19), reasonably close. At least close enough to use S_1 as a first approximation of the maximum salinity of the upper circulation loop (Fig. 6).

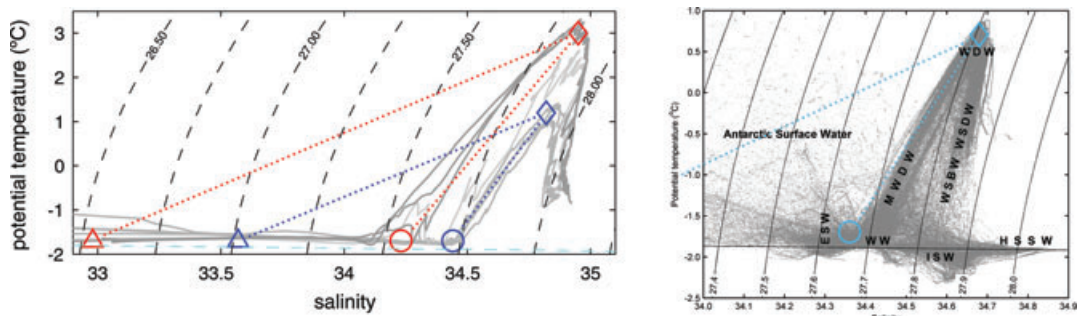


Fig. 6. When the upper layer is formed by melting of sea ice on warmer water during cooling conditions and the energy for mixing is supplied by the wind, the salinity of the upper layer at freezing point is related to the temperature and salinity of the underlying water. The figure shows TS diagrams from the Fram Strait and the Barents Sea inflow branches (left-hand panel) and the Weddell Sea (right-hand panel). The diamonds indicate the salinity and temperature of the temperature maximum, 34.95 and 3.0 °C for the Fram Strait branch (red), 34.82 and 1.2 °C for the Barents Sea branch (blue), and 34.68 and 0.7 °C for the Weddell Sea (cyan). The circles indicate the upper layer salinity for the case of minimum melt rate (eq. 19) and the squares the case when all heat is used for ice melt (eq. 12). For the Fram Strait branch the salinities are 34.23 and 32.98, for the Barents Sea branch 34.44 and 33.57 and for the Weddell Sea 34.36 and 33.63, respectively. The case with minimum ice melt is more in agreement with observations. The figure for the Weddell Sea is adapted from Nicholls et al. (2009).

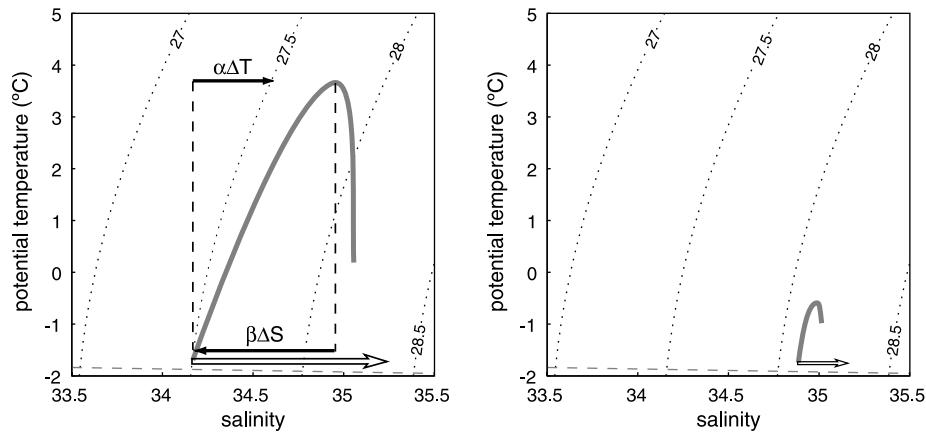


Fig. 7. The density step at the base of the upper layer formed by ice melt is half the density step due to the salinity decrease and equal and opposite to that due to temperature. When the second layer is warm, the density step is large and local freezing and brine rejection cannot break through the interface into the lower layer, increasing its density. However, if the second layer is colder, the salinity of the upper melt water layer becomes higher and the stability of the interface weaker. It is then conceivable that ice formation can increase the salinity in the upper layer sufficiently for the density step to be removed and for saline, convecting parcels to bypass the second water mass and directly enter the deeper layers. The broad arrow indicates θS changes due to freezing.

Once the upper layer is at freezing temperature the only heat available is that entrained from below, which is not enough to supply heat both to ice melt and to the atmosphere. Instead of changing the fraction ϕ_0 we follow Rudels et al. (1999) and assume that the missing heat is provided by formation of new ice. The two processes, entrainment and melting, and ice formation and brine rejection are assumed to operate independently of each other. The fraction of entrained heat going to ice melt is still taken to be ϕ_0 and the brine rejection and haline convection caused by the ice formation is assumed to have only a minor influence on the entrainment rate, indicated by the small value of ε in eq. 15, increasing it slightly (e.g. Stigebrandt, 1981a; Rudels et al., 1999).

Up to the moment when freezing temperature is reached, the ratio $\alpha\Delta T/\beta\Delta S$ is $1/2$, as can be seen if ϕ_0 is inserted into the denominator of the first term in eq. (15), and the stabilizing effect of the salt stratification is twice the destabilizing effect of the temperature stratification (Fig. 7). As the ice formation starts the salinity of the upper layer will increase, and the stability at the interface weakens. If the temperature of the advected water is high when it first encounters sea ice, as is the case both for the Fram Strait branch and for the Barents Sea branch, the possibility to have a salinity increase by brine release in the upper layer sufficient for haline convection to break through the interface and penetrate into the underlying water column is remote. To have communication between the upper layer and the deeper part of the Arctic Ocean, the ice formation has to occur over the shelves, where saline water can accumulate at the bottom and the bottom salinity can increase during winter. As the shelf bottom water eventually crosses the shelf break, it has a density high enough to sink deep into the water column before it reaches its equilibrium density, in spite of entrainment

of less dense ambient water during the descent. Dense water from Storfjorden (e.g. Fig. 5) has been observed at 2000 m in Fram Strait (Quadfasel et al., 1988).

The North Atlantic extends so far north that the cooling of the northward flowing Atlantic water taking place in the Norwegian Sea and the Barents Sea is sufficient for the Atlantic water to reach the density of the Greenland–Scotland overflow water before it encounters sea ice. The interactions with sea ice thus move a part of the Atlantic water from the overflow loop into the surface loop. The waters of the dense loop will be transformed somewhat in the Arctic Ocean by dense boundary plumes that add brine enriched water from the shelves and redistribute entrained water towards deeper levels. The bulk of the overflow water is, however, created already in the Norwegian Sea and the Barents Sea.

Only if sea ice interacts with water initially close to freezing temperature would it be possible for an initial ice melt water layer to attain high enough salinity by ice formation and brine rejection to generate deep reaching open ocean haline convection. Indications that such events can occur were found in the central Greenland Sea in the late 1980s, where in winter 1988 a convective event was observed reaching 1250 m depth and subsequently covered by a less saline and colder surface layer, probably caused by ice melt as warmer water was brought to the surface after the convection (Rudels et al., 1989). Profiles obtained at the same location the following summer indicated that further convection events had occurred and the convection finally reached 2500 m (Fig. 8). The surface layer observed in winter could not have attained a high enough density to convect into the deep by just cooling to the freezing point and additional ice formation was necessary for the observed deep convection. This interpretation assumes that the two

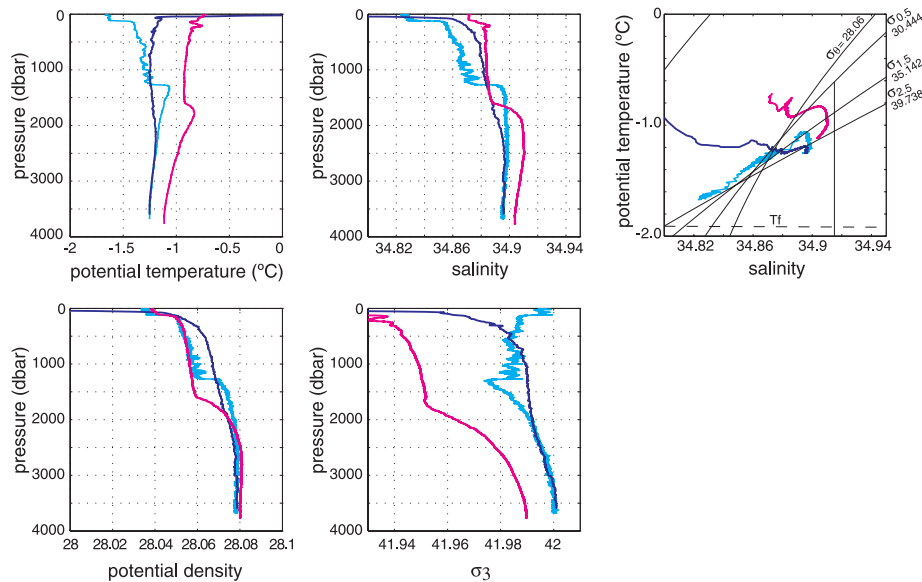


Fig. 8. Stations from the central Greenland Sea taken 1988 winter (cyan) and summer (blue) and in 2002 (magenta), showing the changes in the deep convection. The stations from 1988 indicate an overturning down to 1250 m in mid-winter and a further deepening to 2500 m in late winter. The low salinity surface layer observed in mid-winter 1988 could not have reached a density high enough to convect by cooling alone (see θS diagram). Ice formation and brine rejection would also be necessary. In 2002 the upper layer was too thick and too warm to allow for ice formation during winter and only intermediate thermal convection occurred.

observations really show a time evolution of the same water column, which cannot be proven. In this context it should also be noted that the density anomaly created by the freshwater removal by freezing is significantly ($\sim 10\times$) larger than the density anomaly caused by the corresponding cooling at temperatures close to freezing. The possibility for convective plumes created by brine rejection to bypass the intermediate layers is thus higher than for plumes driven by cooling.

The Greenland Sea has been considered the central area for deep convection and deep-water renewal in the Arctic Mediterranean since the days of Nansen and Helland-Hansen (Nansen, 1906; Helland-Hansen and Nansen, 1909). They postulated that the cooling in winter allows the weakly stratified water column to overturn, occasionally to the bottom. An important factor in this overturning is the preconditioning of the density field, bringing the denser, deeper water as a dome close to the sea surface, leaving only a small volume to be cooled before deep convection can commence. The mechanism behind the doming has commonly been considered to be the dominantly cyclonic wind field present over the Greenland Sea.

In recent years no local deep convection and no local deep-water renewal have taken place in the Greenland Sea. The convection has been limited to the upper 1000–1800 m with localized vortex extending down to 2400 m (Gascard et al., 2002). The doming has disappeared and instead a thick ‘bowl’ of intermediate water has formed above a mid-depth temperature maximum that ultimately derives from the Canadian Basin of the Arctic Ocean (e.g. Rudels 1995). The deeper layers, except for the

small, gradually eroded, remnants of the last local convection residing at the bottom, are renewed advectively from the Arctic Ocean (Meincke et al., 1997). The change in hydrographic structure of the Greenland Sea could partly be connected to the large ice export from the Arctic Ocean occurring in 1996–1997, which lead to ice formation and convection down to 1200 m in the Greenland Sea (Watson et al., 1999). This convection event created a thick pool of Arctic Intermediate Water with enough heat content not to be cooled to freezing temperature in the following winters and no surface water dense enough to penetrate through the intermediate water layer and the mid-depth temperature maximum has been created. Instead the thickness of the intermediate layer has gradually increased, changing the upward doming of the isopycnals to isopycnals pushed downward in the centre of the Greenland Sea (Budéus et al., 1998). Since the wind fields have not changed significantly, this suggests that the doming observed earlier in the Greenland Sea could have been due to preconditioning by previous deep convection, creating the dense doming deep-water mass, rather than by the cyclonic wind field.

The density of the Arctic Intermediate Water formed in the Greenland Sea is nevertheless high enough for it to barely pass over the sill in Denmark Strait. Even if the convection does not reach to the bottom, the Greenland Sea presently creates water that directly contributes to the overflow, perhaps even in larger quantities than during periods of convection extending to the bottom, and the density of the convecting water was too high to allow it to cross the sills at the Greenland–Scotland Ridge.

5 Summary

5.1. Freshwater content

The amount of liquid freshwater in the Arctic Ocean is stable and only responds to changes in freshwater input, either from a reduced ice formation or by increased run-off or precipitation. Perturbations of the volume exchanges with the world ocean, causing changes in the freshwater outflow, lead to increase or decrease in the baroclinic freshwater outflow, which restores the freshwater balance within a couple of years. This dependence on the freshwater content is well known in fjord dynamics (see e.g. Stigebrandt, 1981b) and a discussion along similar lines as here but in another context has been presented in Jakobsson et al. (2007).

5.2. Vertical heat flux

The upward heat flux from the Atlantic layer depends upon the strength of the mechanical mixing and upon the cooling rate, and the heat lost by the Atlantic water is distributed both to ice melt and transferred to the atmosphere. The rate of ice melt depends not only upon the transported oceanic heat but also upon the temperature of the Atlantic water—higher temperatures lead to larger fraction of the heat loss going to ice melt. This creates stronger stability at the interface between the upper layer and the Atlantic water, which reduces the vertical heat flux and the potential for melting ice in the interior of the Arctic Ocean. This then counteracts the effects of warmer Atlantic water entering the Arctic Ocean. Since the largest part of the oceanic heat loss goes to the atmosphere, not to ice melt, the potential for the heat reservoir in the Arctic Ocean to melt sea ice is considerably smaller than the direct estimate of its heat content indicates.

5.3 Double estuary

The combined effects of cooling and freshwater input create a double estuary, capable of sustaining both an estuarine and a mediterranean circulation. The characteristic upper layer salinity, separating the two loops, is determined at the encounter between the northward flowing Atlantic water and sea ice in the marginal ice zone—the warmer the Atlantic water, the lower the salinity. The water supplying the dense water and the overflow loop is mainly formed by cooling before sea ice is encountered. A part of the dense water enters the Arctic Ocean, where it is augmented by dense water formed by freezing and brine rejection on the shallow shelves and sinking down along the continental slope as entraining boundary plumes. Only when strongly cooled water interacts with sea ice will the initial salinity in the upper layer be high enough for subsequent freezing and brine rejection to increase the upper layer salinity and density sufficiently to allow it to convect into the deep. The upper layer is replaced by warmer water from below leading to ice melt and

restratification. When the new upper layer is cooled to freezing temperature, ice formation starts and its density increases until a new convection event takes place. The convection, in spite of being generated by brine rejection, transports freshwater into the deep and eventually there is not enough freshwater to form ice before the stability is removed and the upper layer convects. The subsequent convection will then be thermal, not haline. This leads to a gradual, but slow, deepening of the mixed layer rather than the by-passing convection caused by the larger density anomalies due to freezing. This suggests that the renewal of bottom water in the Greenland Sea by local convection could be the result of a delicate balance of moderate freshwater input at the surface, freezing and weak underlying stratification. These conditions do not appear to be present today.

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References

- Budéus, G., Schneider, W. and Krause, G. 1998. Winter convective events and bottom water warming in the Greenland Sea. *J. Geophys. Res.* **103**, 18513–18527.
- Carmack, E. C., Macdonald, R. W., Perkin, R. G., McLaughlin, F. A. and Pearson, R. J. 1995. Evidence for warming of Atlantic Water in the southern Canadian Basin of the Arctic Ocean: results from the Larsen-93 expedition. *Geophys. Res. Lett.* **22**, 1061–1064.
- Carmack, E. C., McLaughlin, F. A., Yamamoto-Kawai, M., Itoh, M., Shimada, K. and co-authors. 2008. Freshwater storage in the northern ocean and the special role of the Beaufort Gyre. In: *Arctic-Subarctic Ocean Fluxes* (eds R. R. Dickson, J. Meincke and P. Rhines), Springer, Dordrecht, 145–169.
- Coachman, L. K. and Aagaard, K. 1988. Transports through Bering Strait: annual and interannual variability. *J. Geophys. Res.* **93**, 15535–15539.
- Coachman, L. K. and Barnes, C. A. 1961. The contribution of Bering Sea water to the Arctic Ocean. *Arctic* **14**, 147–161.
- Defant, A. 1961. *Physical Oceanography*, Volume I. Pergamon Press, Oxford, 729.
- Dickson, R. R., Rudels, B., Dye, S., Karcher, M., Meincke, J. and co-authors. 2007. Current Estimates of freshwater flux through the Arctic and Subarctic seas. *Progr. Oceanogr.* **73**, 210–230, doi:10.1016/j.pocan.2006.12.003.
- Gascard, J.-C., Watson, A. J., Messias, M.-J., Olsson, K. A., Johannessen, T. and co-authors. 2002. Long-lived vortices as a mode of deep ventilation in the Greenland Sea. *Nature* **416**, 525–527.
- Häkkinen, S. and Proshutinsky, A. 2004. Freshwater content variability in the Arctic Ocean. *J. Geophys. Res.* **109**, C03051, doi:10.1029/2003JC001940
- Hansen, B. and Østerhus, S. 2000. North Atlantic–Nordic Seas exchanges. *Progr. Oceanogr.* **45**, 109–208.

- Helland-Hansen, B. and Nansen, F. 1909. The Norwegian Sea. Its physical oceanography based upon the Norwegian researches 1900–1904. Rep. on Norwegian Fishery and Marine Investigations **II**(1), Det Mallingske Bogtrykkeri, Kristiania, Norway, 390.
- Hurrell, J. W. 1995. Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation. *Science* **269**, 677–679.
- Jakobsson, M., Backman, J., Rudels, B., Nycander, J., Frank, M. and co-authors. 2007. The early miocene onset of a ventilated circulation regime in the Arctic Ocean. *Nature* **447**, 986–990, doi:10.38/nature05924
- Jakobsson, M., Macnab, R., Mayer, L., Anderson, R., Edwards, M. and co-authors. 2008. An improved bathymetric portrayal of the Arctic Ocean. Implication for ocean modelling and geological, geophysical and oceanographic analyses. *Geophys. Res. Lett.* **35**, L07602, doi:10.1029/2008GL033520.
- Kato, H. and Phillips, O. M. 1969. On the penetration of a turbulent layer into a stratified fluid. *J. Fluid Mech.* **37**, 643–655.
- Meincke, J., Rudels, B. and Friedrich, H. 1997. The Arctic Ocean–Nordic Seas Thermohaline System. *ICES J. Mar. Sci.* **54**, 283–299.
- Nansen, F. 1902. Oceanography of the North Polar Basin. The Norwegian North Polar Expedition 1893–96. *Sci. Results* **2**(9), 427.
- Nansen, F. 1906. Northern Waters. Captain Roald Amundsen's oceanographic observations in the Arctic seas in 1901. *Vid-selskap Skrifter I, Mat.-Naturv. kl.* **1**(3). Jacob Dybvad, Christiania, Norway, 145.
- Nicholls, K. W., Østerhus, S., Makinson, K. Gammelsrød, T. and Fahrbach, E. 2009. Ice-ocean processes over the continental shelf of the southern Weddell Sea, Antarctica: a review. *Rev. Geophys.* **47**, RG3003, doi:10.1029/2007RG000250.
- Peixoto, J. P. and Oort, A. H. 1992. *Physics of Climate*. American Institute of Physics, New York, 520.
- Polyakov, I. V., Beszczynska, A., Carmack, E. C., Dmitrenko, I. A., Fahrbach, E. and co-authors. 2005. One more step toward a warmer Arctic. *Geophys. Res. Lett.* **32**, L17605, doi:10.1029/2005GL023740.
- Prinsenberg, S., Hamilton, J., Peterson, I. and Pettipas, R. 2009. Observing and interpreting the seasonal variability of the oceanographic fluxes passing through Lancaster Sound of the Canadian Arctic Archipelago. In: *Influence of Climate Change on the Changing Arctic and Sub-Arctic Conditions* (eds J. C. J. Nihoul and A. G. Kostianoy), Springer, Dordrecht, The Netherlands, 125–143.
- Quadfasel, D., Rudels, B. and Kurz, K. 1988. Outflow of dense water from a Svalbard fjord into the Fram Strait. *Deep-Sea Res.* **35**, 1143–1150.
- Quadfasel, D., Sy, A., Wells, D. and Tunik, A. 1991. Warming in the Arctic. *Nature* **350**, 385.
- Rothrock, D. A., Yu, Y. and Maykut, G. A. 1999. Thinning of the Arctic Sea-Ice Cover. *Geophys. Res. Lett.* **26**, 3469–3472.
- Rudels, B. 1986. The outflow of Polar Water through the Arctic Archipelago and the oceanographic conditions in Baffin Bay. *Polar Res.* **4**, 161–180.
- Rudels, B. 1995. The thermohaline circulation of the Arctic Ocean and the Greenland Sea. *Phil. Trans. Roy. Soc. Lond. A* **352**, 287–299.
- Rudels, B., Quadfasel, D., Friedrich, H. and Houssais, M.-N. 1989. Greenland Sea convection in the winter of 1987–1988. *J. Geophys. Res.* **94**, 3223–3227.
- Rudels, B., Friedrich, H. J., Hainbucher, D. and Lohmann, G. 1999. On the parameterisation of oceanic sensible heat loss to the atmosphere and to ice in an ice-covered mixed layer in winter. *Deep-Sea Res. Part II* **46**, 1385–1425.
- Rudels, B., Marnela, M. and Eriksson, P. 2008. Constraints on estimating mass, heat and freshwater transports in the Arctic Ocean—an exercise. In: *Arctic–Subarctic Ocean Fluxes* (eds R. R. Dickson, J. Meincke and P. Rhines), Springer, Dordrecht, 315–341.
- Serreze, M. C., Barrett, A., Slater, A. J., Woodgate, R. A., Aagaard, K. and co-authors. 2006. The large-scale freshwater cycle of the Arctic. *J. Geophys. Res.* **111**, C11010, doi:10.1029/2005JC003424.
- Shimada, K., Kamoshida, T., Itoh, M., Nishino, S., Carmack, E. and co-authors. 2006. Pacific Ocean inflow: influence on catastrophic reduction of sea ice cover in the Arctic Ocean. *Geophys. Res. Lett.* **33**, L08605, doi:10.1029/2005GL025624.
- Steele, M. and Boyd, T. 1998. Retreat of the cold halocline layer in the Arctic Ocean. *J. Geophys. Res.* **103**, 10419–10435.
- Stigebrandt, A. 1981a. A model for the thickness and salinity of the upper layers of the Arctic Ocean and the relation between the ice thickness and some external parameters. *J. Phys. Oceanogr.* **11**, 1407–1422.
- Stigebrandt, A. 1981b. A mechanism governing the estuarine circulation in deep, strongly stratified fjords. *Estuarine, Coastal Shelf Sci.* **13**, 197–211.
- Stigebrandt, A. 1984. The North Pacific: a global-scale estuary. *J. Phys. Oceanogr.* **14**, 464–470.
- Watson, A. J., Messias, M.-J., Fogelqvist, E., Van Scoy, K. A., Johannessen, T. and co-authors. 1999. Mixing and convection in the Greenland Sea from a tracer release experiment. *Nature* **401**, 902–904.
- Werenskiold, W. 1935. Coastal currents. *Geophys. Publ.* **10**(13), Oslo, Norway, 14.