Stratification and water mass formation in the Arctic Ocean: some implications for the nutrient distribution

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The mixing processes and the water formations (transformations) in the Arctic Ocean are reviewed and their influence on the stratification discussed. The relations between the stratification and the nutrient distribution are examined. The interactions between drifting sea ice and advected warmer and nutrient-rich waters favour an early biological activity. By contrast, in the central Arctic Ocean and over comparably deep shelf areas such as the northern Barents Sea, the possibilities for large productivity are more limited because of late melting, less nutrient supply, and in the central Arctic, less available light. The sedimentation of organic matter on the shelves and the remineralisation into cold, dense waters formed by brine rejection and draining off the shelves lead to a loss of nutrients to the deep waters, which must be compensated for by advection of nutrient rich waters to the Arctic Ocean.

Possible effects of a reduction of the river run-off on the stratification and the nutrient distribution are discussed.

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Introduction

The Arctic Ocean is an enclosed sea with four restricted openings: the narrow and shallow (50 m) Bering Strait communicates with the Pacific Ocean, while the Barents Sea, the channels in the Canadian Arctic Archipelago, and the Fram Strait open to the North Atlantic. The Fram Strait, with a sill depth of 2600 m, is the only deep passage. The depths of the other openings are 200–300 m.

The bathymetry is characterised by the wide and shallow shelf seas north of the Eurasian continent. The mean depth of the Laptev and East Siberian Seas is between 30 and 50 m, and the average depth of the Chukchi Sea is ~50 m. The bathymetry of the Barents and Kara Seas is irregular with several shallow banks and deeper depressions; the depth reaches 400–600 m. The shelf areas north of the American continent are, by contrast, much narrower. The remaining two-thirds of the Arctic Ocean consists of two major basins, the 3000–4000 m deep Canadian and the 4000–5000 m deep Eurasian basin. They are separated by the 1300 m deep Lomonosov Ridge, which runs close to the pole from Siberia to Greenland (Fig. 1).

The climate is severe and the central Arctic Ocean is permanently covered by sea ice. However, because of the seasonally varying radiation balance, a freezing and melting cycle is present. The waters in the Arctic Ocean are formed from the interactions, driven mainly by the freezing and melting processes on the shelves between the river run-off and waters advected into the Arctic Ocean.

About 0.8 Sv of Pacific Water of salinity 32.5 psu enters through the Bering Strait (Coachman & Aagaard 1988). The inflow of the Atlantic (and coastal) Water over the Barents Sea is 1.5–2 Sv with a salinity close to 35 psu (Blindheim 1989; Timofeyev 1963). Most of the Atlantic Water of the West Spitsbergen current recirculates in the Fram Strait; however, about 1 Sv enters the Arctic Ocean (Rudels 1987). Only in the Fram Strait are the in- and outflow of comparable strength; this is the only passage which permits deep water exchange.

The waters formed or transformed in the Arctic Ocean are (Fig. 2):

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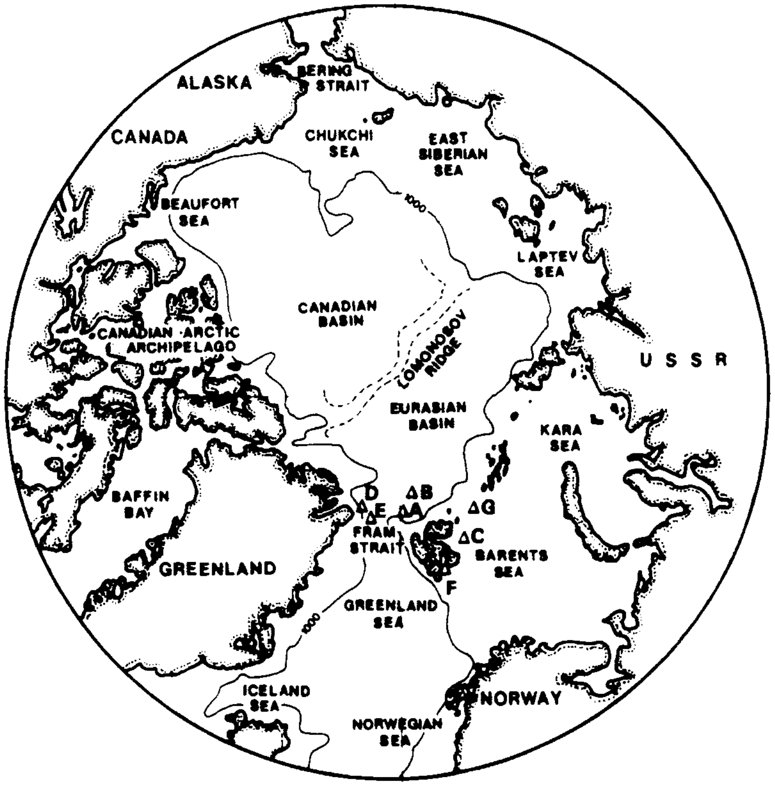


Fig. 1. Bathymetric map of the Arctic Ocean (from Aagaard & Carmack 1989). The positions of the hydrographic stations discussed in the text are indicated.

1. The 50 m deep Polar Mixed Layer, with freezing temperature and salinity of about 32.7 psu close to the Fram Strait and somewhat lower in the Beaufort Sea.

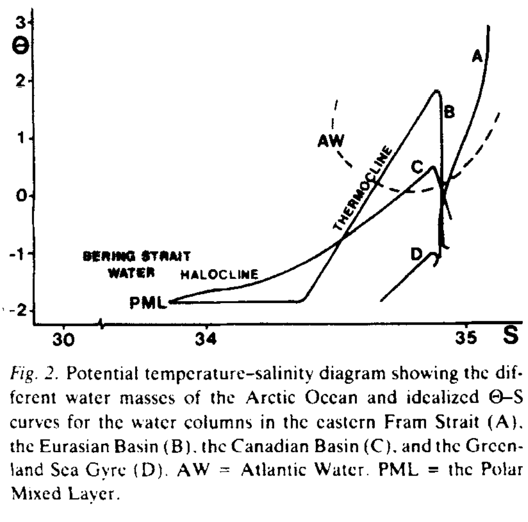


Fig. 2. Potential temperature–salinity diagram showing the different water masses of the Arctic Ocean and idealized Θ–S curves for the water columns in the eastern Fram Strait (A), the Eurasian Basin (B), the Canadian Basin (C), and the Greenland Sea Gyre (D). AW = Atlantic Water. PML = the Polar Mixed Layer.

2. The halocline between 50 m and 250 m with salinities ranging from below 33 to 34.4 psu. The temperature is mostly close to freezing but increases at the lower boundary to 0°C. A temperature minimum is usually found in the halocline.
3. The 400–600 m thick layer with temperatures above 0°C, which is defined as the Atlantic layer. Its salinity increases with depth from 34.4 to 34.9 psu. The transition layer between the halocline and the temperature maximum at the Atlantic layer is denoted the thermocline.
4. Deep Waters below 800–1000 m with salinities of 34.93–34.95 and potential temperatures ranging from 0°C to −0.95°C at the bottom. The Canadian Basin Deep Water is warmer, about −0.5°C, because the Lomonosov Ridge prevents the coldest and densest deep water entering through the Fram Strait from reaching the Canadian Basin. This temperature difference was in fact one of the clues leading to the discovery of the Lomonosov Ridge (Worthington 1953).

About 1 Sv of Polar Mixed Layer and halocline waters exits through the Canadian Archipelago (Rudels 1986a). The outflow of the rest of the upper waters (1 Sv) and of the Atlantic and deep layers occurs through the Fram Strait. The strength of the deep water exchange is not well known, but probably 1 Sv enters and 1.5 Sv leaves the Arctic Ocean (Rudels 1986b). The outflow of Atlantic Water through the Fram Strait is about 0.7 Sv (Rudels 1987).

The fresh water supply from the river run-off is large (0.1 Sv), but a comparable export of sea ice takes place through the Fram Strait (Aagaard & Carmack 1989). However, some fresh water is added to the water column and contributes to the strong stability of the Arctic Ocean.

Below we will discuss some aspects of the thermodynamically driven mixing processes in the Arctic Ocean and their influences on the water mass formations and the nutrient distribution. For a more thorough discussion of the oceanography of the Arctic Ocean we refer to Coachman & Aagaard (1974) and to the Atlas of the Arctic Ocean edited by Gorshkov (1980).

The stratification of the Arctic Ocean

In contrast to lower latitudes, the local influences on stratification and water mass formation in the Polar Seas are dominated by freezing and melting rather than by cooling and heating. The melting sea ice keeps the sea surface temperature close to freezing even in summer and creates a low-salinity surface layer with strongly reduced density.

Because sea ice moves relative to the water, a low-density surface layer is created in the frontal areas by ice drifting over and melting on warmer waters without heat being added at the sea surface. Fig. 3 shows such a situation at a hydrographic station north of Svalbard, where warm Atlantic Water encounters ice and cold polar waters. The positions of the different stations are indicated in Fig. 1.

When the ice melt is due solely to increased solar radiation, the salinity and density decrease are less, and the ice melting is slower and happens later in the season. This situation occurs when the ice rests on top of a cold layer previously homogenised by freezing and convection; it can be seen in the profiles from a station taken further north into the Arctic Ocean (Fig. 4). The heat stored in the Atlantic Water is isolated from the sea-surface by a 80 m deep convection layer and cannot influence the ice melt.

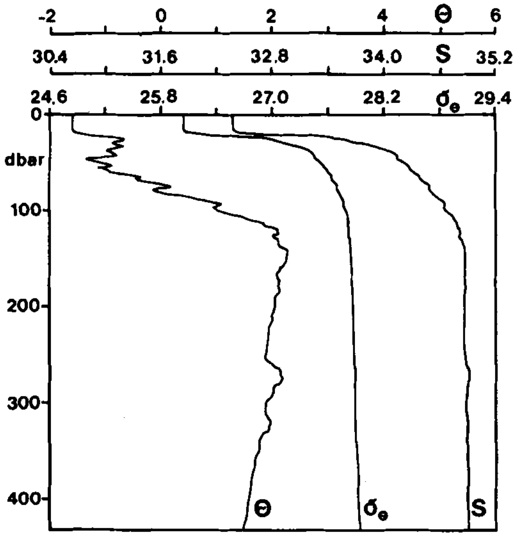


Fig. 3. Potential temperature, salinity, and density profiles for station A in the frontal zone north of Svalbard showing ice melting on warm Atlantic Water.

During the cooling phase, the fresh water added by melting is again removed as newly formed sea ice. Brine is released and the upper layers are homogenised. Depending on the location, the

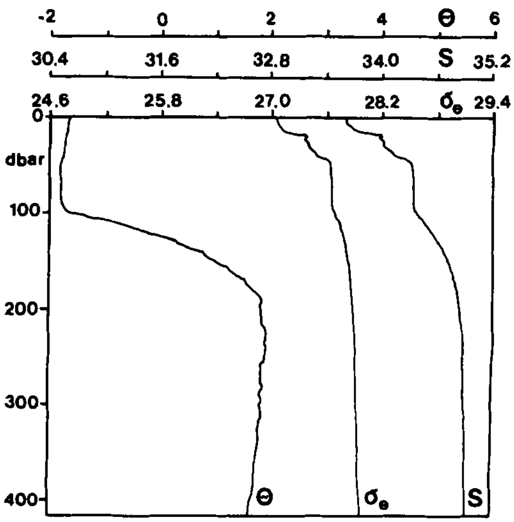


Fig. 4. Potential temperature, salinity, and density profiles for station B in the Sofia Deep north of Svalbard. The ice melt is due to solar radiation.

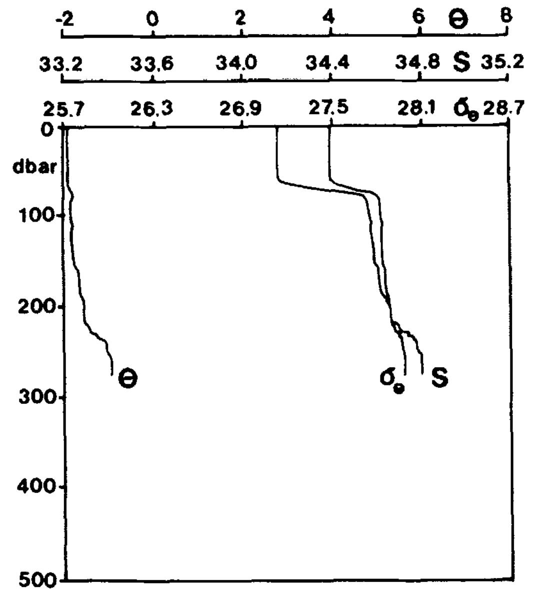


Fig. 5. Potential temperature, salinity, and density profiles for station C in the northern Barents Sea taken in October during the freezing phase.

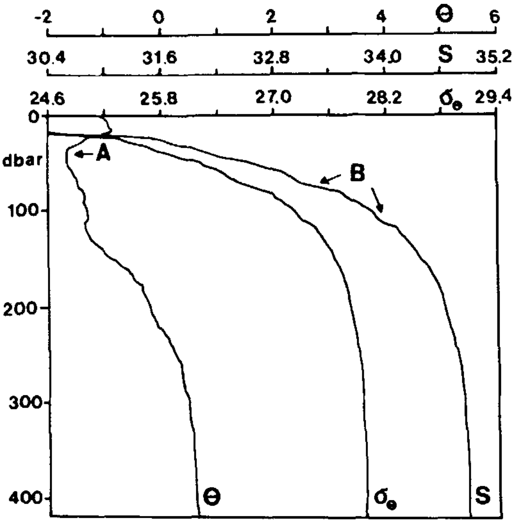


Fig. 6. Potential temperature, salinity, and density profiles for station D in the area of the Arctic outflow in the western Fram Strait. A = the Polar Mixed Layer. B = the halocline.

convection reaches down to the halocline, or to the Atlantic layer, or on the shelves perhaps down to the bottom. The profiles shown in Fig. 5 were taken in October in the northern Barents Sea and reveal the gradual reformation of a deep winter convection layer. If another metre of sea ice were formed that winter, the convection would, neglecting advection effects, reach 200 m.

If the convection penetrates to the bottom, as it is likely to do on the shallow parts of the shelves, brine-enriched water accumulates at the bottom and will eventually find its way to the deep basins. However, since the shelves are wide this will take some time. Meanwhile, the salinity increases due to the continued addition of brine-enriched waters throughout the winter. The freezing cycle on the shelves may therefore create waters dense enough to ventilate the entire water column of the deep basins. However, the most important product is the halocline. This water mass is not dense enough to sink into the Atlantic layer, but it is denser than the Polar Mixed Layer and forms an intermediate water mass above the Atlantic Water. Depending upon the local conditions at the different shelf regions, waters at the freezing point but which occupy a large range in salinity (as long as the density is above that of the Atlantic layer) may form and supply the halocline.

Remnants of this feature can be seen in profiles taken in the western Fram Strait (Fig. 6). Below the temperature minimum at 50 m, the salinity increases drastically as compared to the small increase in temperature. This density step limits the convection to the upper 50–60 m and the salinity increase cannot arise from direct mixing between the Polar Mixed Layer and the Atlantic Water (Fig. 2). This was pointed out by Coachman & Barnes (1962). By making the vertical entrainment of Atlantic Water vanishingly small, the presence of the halocline in most of the Arctic Ocean (Coachman & Aagaard 1974) prevents direct communication between the Atlantic Water and the Polar Mixed Layer and the sea ice. It has been suggested that the isolating effect of the halocline is important for retaining the sea-ice cover in the Arctic Ocean (Aagaard & Coachman 1975). This problem will be discussed in connection with the effects of reduced river run-off.

The Θ–S diagram from the stations shown in Figs. 4 and 6 are very similar (Fig. 7A and B). However, at the station north of Svalbard (Fig. 7A), the water distributed along the freezing line is due to summer ice melt and the local convection reaches down to the Atlantic layer. At the other station, a substantial part of the salinity increase occurs below the temperature minimum but above the main thermocline, which indicates the presence of a more saline water mass, with tem-

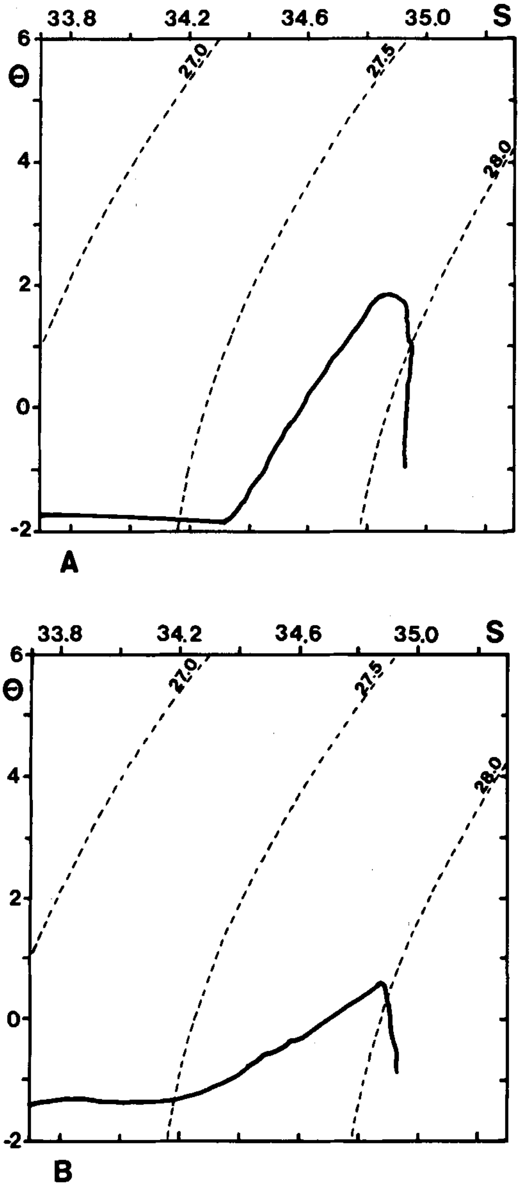


Fig. 7. Potential temperature–salinity diagrams for station B (A) and station D (B) shows indications of the halocline proper.

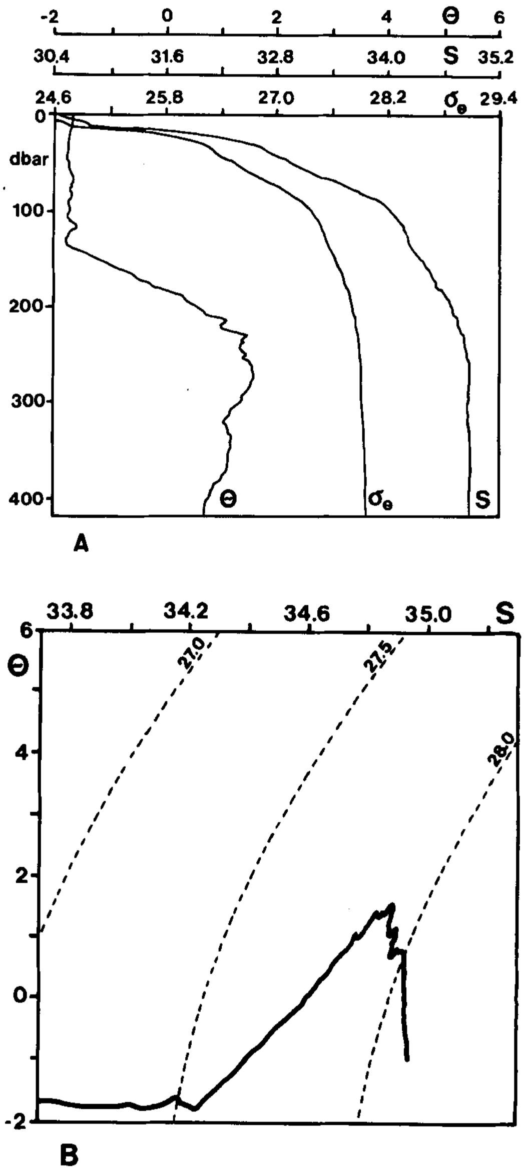


Fig. 8. Potential temperature, salinity, and density profiles (A) and Θ–S diagram (B) for station E in the northern Fram Strait. The station indicates lateral mixing between the Polar outflow and Atlantic water freshened and cooled, perhaps by freezing, close to Fram Strait.

perature close to freezing above the Atlantic layer. This water, while necessarily an advective feature, may not represent the halocline proper. It could also form in the northern vicinity of Fram Strait and intrude laterally beneath the Polar Mixed Layer. At a neighbouring station this appears to be the case (Fig. 8), which shows that it is treacherous to discuss the features of the halocline using data only from the Fram Strait; to distinguish the different water masses, more tracers than temperature and salinity are needed (see below, The nutrient distribution).

The changes in the Θ–S structure of a water column, which enters through the Fram Strait,

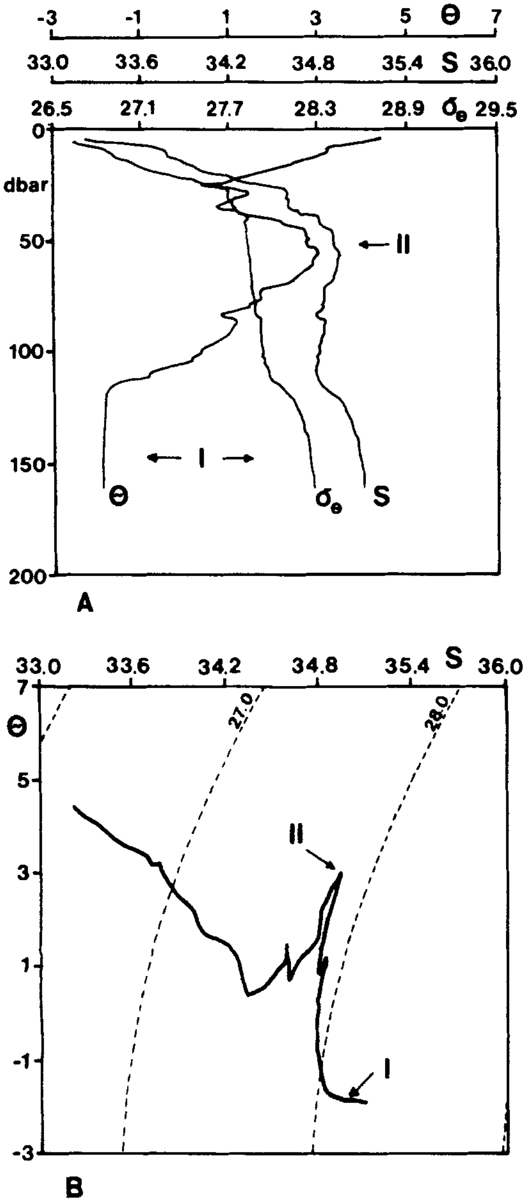


Fig. 9. Hydrographic structure at station F in Storfjorden. A. Vertical profiles of potential temperature, salinity, and potential density. B. Potential temperature-salinity diagram. The water at freezing temperature but with increasing salinity below 120 m (I) indicates accumulation of brine-enriched water. Note the coinciding temperature and salinity maxima (II).

are primarily but not entirely due to heat loss and dilution by river run-off and low salinity Bering Strait inflow. This is seen in Fig. 2, which shows the Θ–S curves for the North Atlantic waters entering through the Fram Strait, the water columns of the Eurasian and Canadian basins of the Arctic Ocean, and the waters of the Greenland Sea. In addition to the cooling and freshening of the upper layers, salinity, temperature and stability increases occur in the deep and bottom layers in the Arctic Ocean. This effect, which is not seen in the Greenland Sea in spite of a comparable heat loss, arises because the salinity of the brine-enriched water on the shelves may become much higher than what is possible in the surface water in the deep, weakly-stratified Greenland Sea Gyre (Rudels 1990). Freezing causes transport of salt into the deep, and when the ice melts, a low density surface layer is formed. The freezing and melting cycle thus adds to the river run-off and the Bering Strait inflow in creating the stable stratification of the Arctic Ocean.

This implies that some parts of the shelves, particularly shallow and/or confined areas, create water which penetrates deep into the Arctic Ocean water column, occasionally reaching the bottom. Such dense waters have been observed in the eastern Barents Sea (Nansen 1906; Midttun 1985) and in Storfjorden (Anderson et al. 1988; Fig. 9). As the shelf water sinks through the water column, it entrains ambient lighter water, and its Θ–S characteristics change. The shelf water, which intrudes in the upper part of the water column, lowers its temperature and salinity, while the denser, deep sinking waters have high initial salinity and will entrain so much Atlantic Water that they make the deeper layers warmer and saltier. Fig. 10 shows a schematic picture of the Θ–S changes of the descending water parcels and of the ambient water column. For details, see Rudels (1986b) and Quadfasel et al. (1988).

As a curiosity we may note the frequent occurrence of coinciding temperature and salinity maxima (Fig. 9). If a water mass with extreme properties lies sandwiched between two waters, the Θ–S curve will, because of vertical mixing, experience a separation of the maxima (minima). This happens regardless of the nature of the vertical mixing process – mechanical mixing, salt fingering, diffusive interfaces (Rudels 1987). Coinciding maxima are therefore a sign of recent intrusive mixing.

In the present case, waters from the shelves, occupying a broad density (salinity) range, penetrate into the water column. The large volume of Atlantic Water is limited to a small density

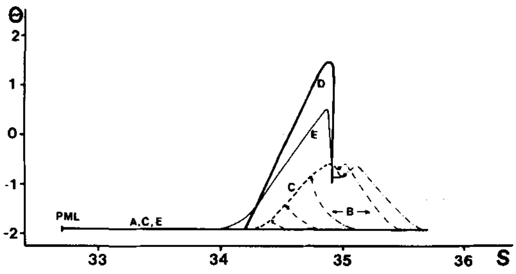


Fig. 10. Potential temperature–salinity diagram showing schematically the cold, dense water formed on the shelves (A), the evolution of the Θ–S characteristics of the shelf water as it sinks as entraining density flow down the continental slope (B), the curve connecting the Θ–S characteristics of the different density flows at their optimal depth (density level) (C), a "young" Arctic water column north of Svalbard unaffected by the intruding waters from the shelves (D), and an "old" Arctic water column transformed by the shelf water as found northeast of Greenland (E). Adapted from Rudels (1986b).

interval and the shelf water will dominate in the density ranges above and below that of the Atlantic Water and also reduce the temperature and salinity more in ranges where the Atlantic Water volume is smaller. The Θ–S value representing the largest initial volume of Atlantic Water will remain, briefly, as an extremum, with coinciding temperature and salinity maxima.

The nutrient distribution

As the water mass formation and stratification in the Arctic Ocean are determined by the interaction between advected waters and the freezing and melting processes, so is the nutrient distribution. Nutrients are added to the Arctic Ocean by inflows of Atlantic and Pacific waters and by the rivers. The Pacific Water contains more nutrients, especially silica, and supplies twice as much silica in spite of its smaller inflow. The nitrate concentrations of the two waters are more equal. The silica concentrations of the rivers is three to four times as large as that of the Pacific Water and the input is comparable to that of the Atlantic Water. (Codispoti & Owens 1975; Anderson & Dryssen 1981; MacDonald et al. 1987).

The most productive areas are the frontal zones, where sea ice drifts over and melts on warmer nutrient rich Atlantic and Pacific waters, or on the shelves, when the ice breaks up and the spring run-off starts. In these cases a shallow strong stratification is created, capable of keeping the organisms trapped in the sunlit, productive zone, allowing them to utilise the nutrient supply in full.

However, in the central Arctic Ocean as well as in the northern Barents Sea, the sea ice floats on water homogenised the previous winter by convection and at freezing temperature. A stable stratification will not form until the solar radiation is intense enough to melt some of the ice; the production will start later. In addition, the presence, also in summer, of sea ice diminishes the amount of light penetrating into the water and reduces the productivity despite the stratification.

If light is sufficient, the plankton bloom rapidly depletes the nutrient reservoir and organic matter has to be decomposed and the nutrients resupplied. Essentially four possibilities exist:

1. The mineralisation occurs continuously in the top low-salinity layer and the nutrients are reused immediately.
2. Organisms fall out of the stable top layer and are dissolved in the cold water below. Nutrients can be mixed into the top layer during the summer by mechanically driven turbulent entrainment. Some assimilation also occurs below the top layer once the ice and the surface bloom are gone and the light can penetrate deeper. However, the main reintroduction of nutrients is due to the convection and redistribution of water masses in the following winter.
3. On the shallow shelves some of the organic matter reaches the bottom before it dissolved. The nutrients will eventually leak from the sediments into the bottom water of the shelf.

In the Chukchi Sea, areas favourable for dense water formation create a dense bottom water, which covers most of the Chukchi Sea. This implies that the winter homogenisation in other parts can only reclaim nutrients which are dissolved above this dense bottom water as in case 2, while the nutrients remineralised from the sediments are trapped in the bottom water.

A similar situation exists in parts of the Kara and Barents Seas. The convection reaches the bottom; by accumulation of brine-enriched water throughout the winter, the bottom waters attain such high densities that they cannot be reintroduced into the photic zone and the nutrients dissolved from the sediments are lost for the production.

In the Laptev and East Siberian Seas a saline, cold bottom water is present which, because of the low salinity of these seas, cannot be formed locally in spite of the shallow depths. The nutrients remineralised from the sediments accumulate in this bottom layer, which in the East Siberian Sea probably is advected from the Chukchi Sea (Codispoti & Richards 1968), and in the Laptev Sea perhaps from the area around Severnaja Zemlja. The nutrients in these seas are then primarily resupplied by the rivers. On the McKenzie shelf the bottom water appears to be locally formed. However, in some areas close to the coast, where the fresh water supply is large, convection to the bottom may be prevented (MacDonald et al. 1987, 1989).

4. In the central Arctic Ocean, north of Svalbard and in the northern Barents Sea, a permanent pycnocline is present below the cold homogenous layer. Organisms falling from the productive zone may sink through the pycnocline into the deep Arctic Ocean or in the northern Barents Sea into the Atlantic Water. The winter convection only homogenise the cold layer in the deeper parts of the northern Barents Sea and the Polar Mixed Layer in the Arctic Ocean; only nutrients dissolved in these layers will be reclaimed for the next season. This leads to low nutrient content and low productivity in the northern Barents Sea despite sufficient light conditions.

The waters from the different shelves and frontal areas eventually enter the central Arctic Ocean. However, the characteristics they have required on the shelves allow them to be identified in the interior of the Arctic Ocean. Concentrating on the upper waters (above the Atlantic layer) we follow Jones & Anderson (1986) and distinguish between the Polar Mixed Layer, the upper halocline, and the lower halocline.

The upper halocline ($S = 33.1$ psu, $\Theta = -1.5$ psu) is characterised by a nutrient maximum, especially conspicuous in the silicate concentration, while the lower halocline ($S = 34.2$ psu, $\Theta = -1.1$) has a low nutrient content and a minimum NO value ($NO = 9NO_3 + O_3$, Broecker 1974). The NO values in the Polar Mixed Layer and the upper halocline are almost the same, the upper halocline value being somewhat higher (Jones & Anderson 1986). NO is a conservative tracer and takes into account the utilisation of oxygen during the remineralisation of organic material.

The waters on the Laptev, East Siberian, and McKenzie shelves have such low salinity throughout the year that they, despite the freezing and convection in winter, only supply the Polar Mixed Layer. This is also true for water leaving the Chukchi shelf in the summer. However, Jones & Anderson (1986) have shown that due to the high initial salinity of the Bering Strait inflow, water formed by brine rejection on the Chukchi shelf acquires a density high enough for it to sink beneath the Polar Mixed Layer and enter at about 100 m depth. While it drains off the shelf, it becomes enriched by nutrients leaking from the sediment and it shows up as the nutrient maximum of the upper halocline. The upper halocline also receives the bottom water advected over the East Siberian Sea and perhaps also that of the Laptev Sea.

The origins of the lower halocline appear to be more difficult to establish. The cold homogenous layers north of Svalbard (Fig. 4) and in the northern Barents Sea (Fig. 11) have approximately the right salinity, low NO values, but perhaps too low silicate values (Anderson & Dryssen 1980; Danielson 1980). The dense shelf water formed in the southern Kara Sea has higher salinities and would enter deeper in the water column. Only on the shelf areas north and east of the Ob and Yenisey rivers are waters with the appropriate Θ–S characteristics observed (Garcia 1969; Milligan 1969). This water also has higher silicate concentrations due to the river run-off.

In a recent paper, Anderson & Jones (1991) identified several branches of low-NO water in the lower halocline density range in the Nansen basin, without observing any signs of the upper halocline water. The observations in the northern Barents Sea and north of Svalbard then suggest that this water mass is primarily formed, not by brine accumulation on the shelves, but in a brine-driven open ocean convection vertically limited by the density gradient of the Atlantic Water below.

The water is supplied by exchanges of Atlantic Water across the fronts. Melting ice preconditions the water column for ice formation (Fig. 3) and a cold, fairly deep homogenous layer is then formed by freezing and brine rejection the following winter (Fig. 4). Part of this water recirculates to the Fram Strait (Figure 8), while part flows into the Nansen basin and further into the

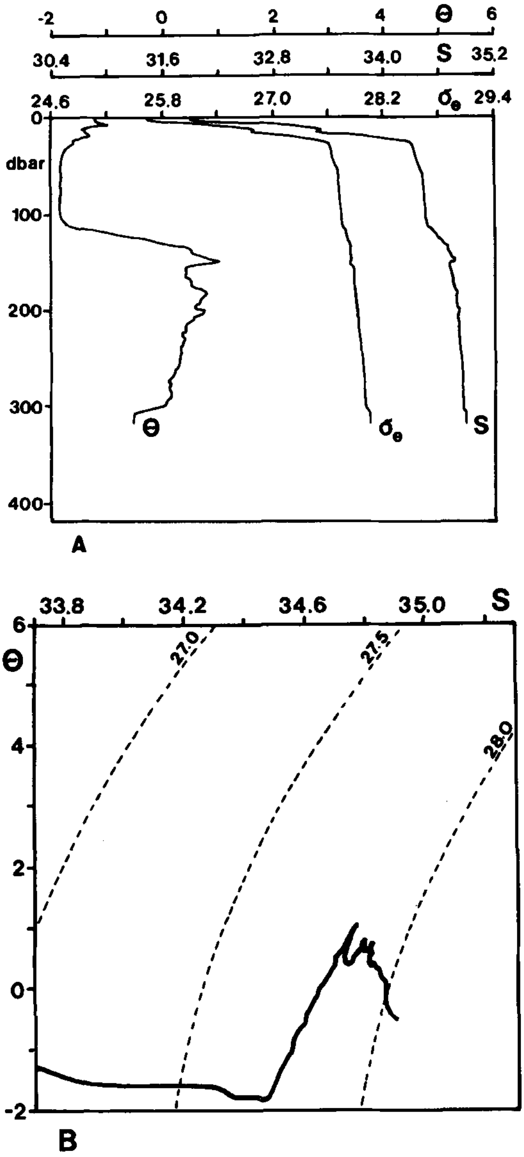


Fig. 11. Potential temperature, salinity, and density profiles (A) and Θ–S diagram for station G in the northern Barents Sea (B).

Arctic Ocean, forming together with the water produced in the northeast of the Kara Sea a barrier between the Polar Mixed Layer and the Atlantic Water. In the Canadian Basin it enters below the upper halocline, and it is still distinguished as a mode water off the McKenzie shelf, indicating that its volume is not negligible (Carmack et al. 1989). We then have the somewhat paradoxical situation that the shelf products created by accumulation of brine are found above the lower halocline formed mainly by open ocean convection. This conjecture is supported by the low nutrient content implying little or no dissolution of organic sediment from the bottom and by the low NO value.

All cold waters in the Arctic Ocean are formed by freezing and their oxygen saturation values should be comparable. Moreover, the shallow photic zone created by the ice melt will be depleted of nutrients during the plankton bloom. However, any supersaturation of oxygen due to biological activity is likely to be lost during the cooling phase when the surface water is brought to the freezing point. When the surface water is mixed down by brine-driven convection in fall and winter, its NO value should be determined by the oxygen saturation value. If the winter convection brings deeper water with partly regenerated nutrients to the surface, the oxygen lost by the respiration can be replaced by contact with the atmosphere and the NO value will increase. This process would be more effective over a shallow shelf where the overturning is more intense than the deeper layers north of Svalbard and in the northern Barents Sea. Compared to the upper halocline, the lower NO value of the Polar Mixed Layer arises because it is supplied not only by the winter waters but also by the summer waters of the Laptev and East Siberian Seas as well as by waters from the McKenzie Shelf, perhaps also by summer waters from the Chukchi Sea. These waters are homogenised in the interior of the Arctic Ocean, but since the summer input is likely to have a low nutrient content, the NO value of the Polar Mixed Layer will be smaller.

The source waters for the lower halocline are initially at freezing temperature and they are heated by the Atlantic layer in the interior of the Arctic Ocean. The most likely mechanisms for the heat flux is double diffusive convection through diffusive interfaces. Because of the small dependence of density on temperature in cold waters, a temperature increase of the halocline by 1°C by double diffusive convection implies a salinity decrease of only 1/100 psu. The relative salinity decrease of the Atlantic layer for a corresponding temperature decrease will be somewhat larger due to the non-linearity of the equation of state.

We also notice from the Θ–S curves (Fig. 7) that the temperature decrease of the Atlantic layer, despite its larger volume, is larger than the corresponding increase in the halocline. This cannot be explained by a heat loss of the halocline to the mixed layer because of the temperature minimum present in the halocline (Treshnikov & Baranov 1972), indicating that an additional cooling mechanism for the Atlantic layer is present. This mechanism is the isopycnal intrusion of cold, dense shelf water, which has been described above (The stratification of the Arctic Ocean) and illustrated in Fig. 10.

The summer and winter situations along a vertical section of the Arctic Ocean from Fram Strait towards Siberia are presented schematically in Fig. 12. The highest biological production occurs in the areas where melting sea ice interacts with entering nutrient-rich waters as in the Chukchi Sea, in the southern Barents Sea, and on the shelves in the vicinity of the rivers. By contrast,

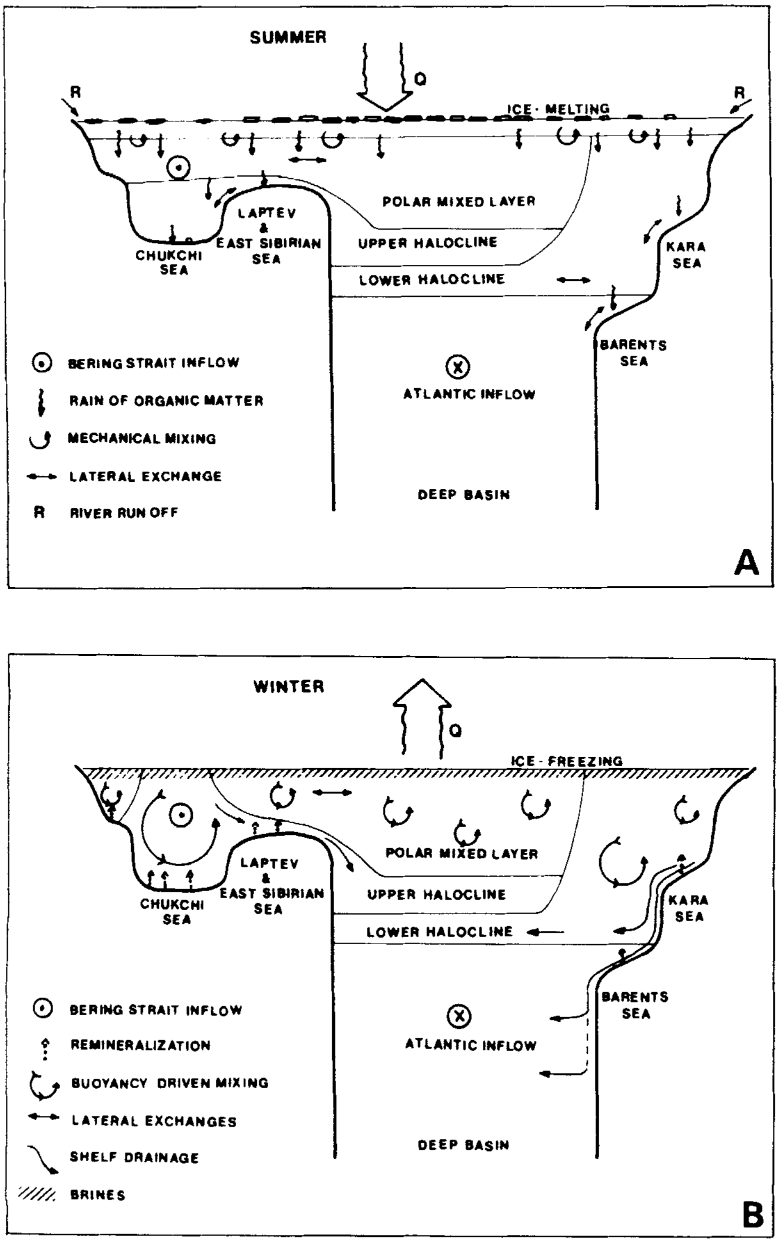


Fig. 12. Schematics showing the summer (A) and the winter stratification (B) of the Arctic Ocean. The view is from Fram Strait towards Siberia. The relative depth of the Polar Mixed Layer is exaggerated. Q = heat exchange. R = river run-off.

because nutrients are only introduced from the cold layer below, the situation in the northern Barents Sea is less favourable. The productivity there observed at the ice edge is likely due to the retreat of the sea ice, opening and stratifying new waters with unused nutrients.

The freezing and melting cycle, which creates the initial stratification necessary for plankton growth, also limits the production by reducing the nutrient reservoir since nutrients dissolved in the dense shelf waters sink into the deeper layers and are lost for the production.

Most of the nutrient-rich upper halocline leaves the Arctic through the Canadian Arctic Archipelago, where high silicate values are observed (Jones & Coote 1980). Anderson & Dyrssen (1981) argued from high silicate observations made in the Fram Strait that some of the upper halocline water also exits through the Fram Strait. However, at the same levels the NO values are as low as 390 (Anderson & Dyrssen 1980). Low NO values are common in the upper layers in the western Fram Strait (Anderson & Jones 1991) and may arise from lateral mixing with freshened, cooled and nutrient-depleted Atlantic Water recirculating in the strait. However, whether or not it is possible to lower the NO value without also removing the silicate maximum is another matter. The Θ–S structures of these stations (Figs. 6 and 8) do not clearly exhibit the upper halocline characteristics, which if present have been considerably diluted by lateral mixing. The deep waters all exit through the Fram Strait and eventually cross the Greenland–Scotland Ridge. The nutrients lost to the system, either by export or by sedimentation, on the shelves or in the deep basins have to be resupplied.

The effects of reduced river run-off

The interactions between the fresh water supply, the freezing and melting cycle, and the shelf areas are clearly shown when a situation with less river run-off is considered. Changes that result in the stratification have been described in some detail by Rudels (1989), and the discussion will concentrate on the possible effects on the nutrient distribution.

The waters on the shelves become more saline. However, because the shallow waters on the shelves will still be brought to freezing point, the ice formation rate on the shelves will remain unchanged. The higher initial salinity will lead to a smaller supply of water to the Polar Mixed Layer and the halocline, and more to the deeper layers. The interaction between Atlantic Water and the Polar Mixed Layer will not, as now, primarily occur on the shelves and at the shelf breaks. Instead, an entrainment of Atlantic Water, through a reduced halocline directly into the Polar Mixed Layer, can take place over most of the Arctic Ocean (Rudels 1989).

The heat thus added to the Polar Mixed Layer reduces the annual net ice production in the central Arctic Ocean and acts to retain a large density gradient between the Atlantic Water and the Polar Mixed Layer, preventing the entrainment from becoming vigorous. The ice cover will be thinner, but not much, and the high albedo should be maintained. The heat loss of the ocean-ice system remains large and the lower run-off would not lead to large climatological changes.

The entrainment makes the salinities of the different Eurasian shelf seas more equal, and the Laptev and East Siberian Seas will produce comparably denser waters. The importance of the Pacific inflow as a fresh water source increases, and water formed by freezing in the Chukchi Sea will be less dense compared to waters from the other shelves than in the present situation.

The water formed in the Kara Sea would attain densities high enough for it to sink into the deep waters. However, a substantial reduction of the run-off would lead to a loss also of the winter waters of the Laptev and the East Siberian Seas to the deep ocean. The strong stability of the Arctic Ocean water column is maintained, but now a greater part of the stratification is due to the formation of dense, deep waters than to

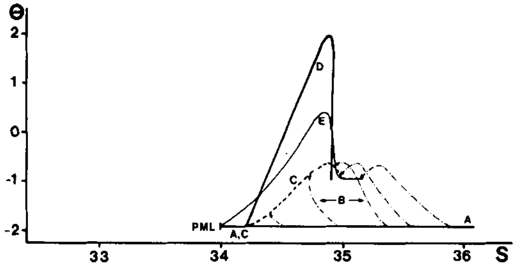
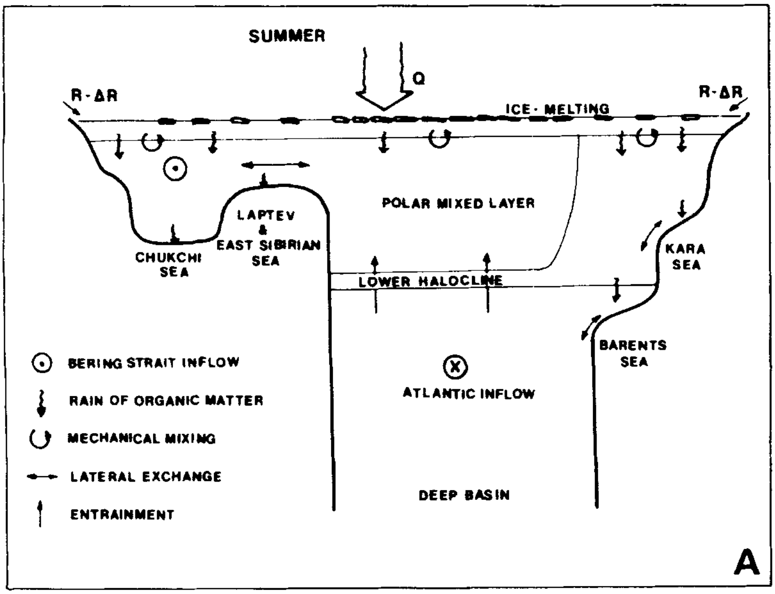


Fig. 13. Potential temperature–salinity diagram showing schematically the interactions between the shelf water and the deep basin water column in the case of reduced river run-off.

the freshening of the upper layers (compare Figs. 10 and 13).

More nutrients will be supplied by entrainment of Atlantic Water, and part of the nutrients of the upper halocline, which at present is lost for the production, might enter the now deeper Polar Mixed Layer. The nutrient input from the rivers will decrease and the situation might arise where the Polar Mixed Layer, enriched with nutrients from entrained Atlantic Water and from the Chukchi shelf, supplies the nutrients for the summer bloom in the Kara, Laptev, and East Siberian Seas. The organic matter then falls to the bottom, nutrients are remineralised and leak into the dense winter water to sink into the deep ocean.

The northern Barents Sea would still be an area where convection will not reach the bottom and no nutrients can be resupplied from the



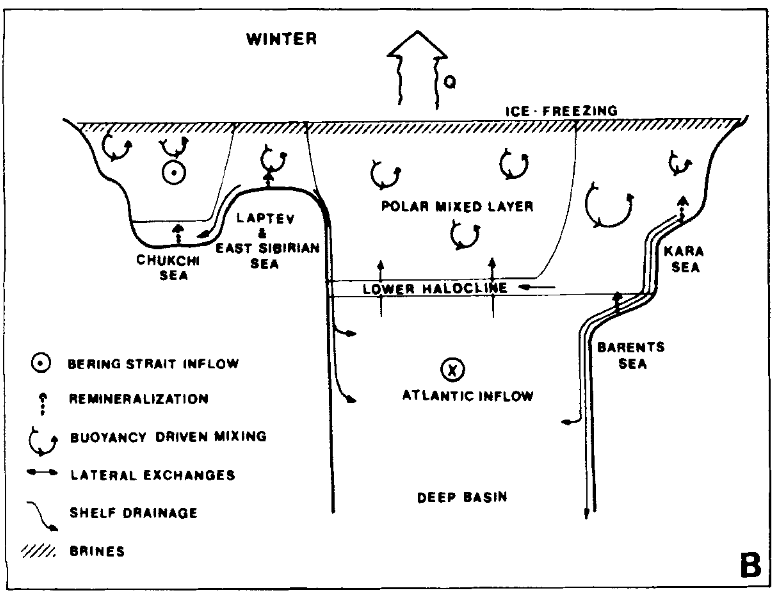


Fig. 14. Schematics showing possible changes in stratification and mixing when the river run-off is reduced by ΔR.

sediments. It remains a source, together with the area north of Svalbard, for a much reduced "lower" halocline, and its nutrient reservoir must, as today, be replenished by a flux of Atlantic Water across the fronts north of Svalbard and in the Barents Sea. The nutrient distributions in this imagined scenario are schematically shown in Fig. 14.

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