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Double-Diffusive Convection and Interleaving in the Arctic Ocean – Distribution and Importance

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Abstract

Beneath its ice cover the Arctic Ocean is a low energy environment. The weak turbulent activity allows other, more esoteric mixing mechanisms to become important in transforming the water masses. One such process is double-diffusive convection, which is triggered by the different molecular diffusion rates of heat and salt and utilises the potential energy stored in the unstably stratified component, heat or salt, to increase the vertical transports. Cold, fresh water above warm, saline water leads to the formation of diffusive interfaces, while warm and saline water above cooler and fresher water results in saltfingers. The former situation is more representative of the Arctic Ocean and the vertical heat transport through diffusive interfaces could contribute significantly to the upward flux of heat from the subsurface warm Atlantic water to the upper layers. The lateral property contrasts between the different inflow branches to the Arctic Ocean and between the boundary current and the water columns of the different basins allow finite lateral disturbances to create intrusions and inversions in the temperature and salinity profiles. These in turn cause double-diffusive transports, which generate convergences and divergences in the vertical buoyancy fluxes, reinforcing the lateral interleaving between the different water masses. Although the interleaving structures are most prominent at frontal zones, they appear to extend over entire basins and are encountered in almost any depth range and in all types of background stratification, saltfinger unstable, diffusively unstable, and when both components stably stratified. The interleaving could be a mechanism that redistributes heat and salt from boundary current at the rim to the interior of the deep basins. However, the formation process(es) for the interleaving layers has (have) not yet been adequately described, and their importance for lateral mixing have not been determined. Here the main features and the distribution of the intrusions in the Arctic Ocean are presented, different explanations for the large extent of the interleaving layers are discussed, and their possible significance for the water mass transformations is evaluated.

Key words: Double-diffusive convection, intrusions, double-diffusive interleaving, saltfingers, diffusive interfaces, Arctic Ocean, Arctic Ocean circulation, Arctic Ocean heat exchange

1. Introduction

The Arctic Ocean, because of the strong stability in its upper layer and its sea ice cover, is a low energy environment (*Padman*, 1995). This in spite of the forcing provided by the cooling during winter. The atmospheric meridional transport of water vapour leads to a large river runoff to the almost enclosed Arctic Ocean (Fig. 1). The net precipitation over the Arctic Ocean itself is also substantial and additional freshwater is carried by the comparatively low salinity Pacific water entering the Arctic Ocean through Bering Strait (*Serreze et al.*, 2006). Furthermore, the strong positive radiation balance in summer promotes a large seasonal ice melt and much of the heat loss in winter is used to reform, and slightly increase, the ice cover. The excess ice is eventually exported, mainly through Fram Strait to the Nordic Seas and to the North Atlantic. The stratification and the ice cover reduce the wind generated turbulence and mixing mechanisms other than those mechanically driven become more important, either in their own right or in combination with mechanical processes like the generation and breaking of internal tides. The internal wave energy in the Arctic Ocean is generally quite low, probably due to the presence of the sea ice cover. However, this may lead to the enhancement of some internal wave motions (*Padman*, 1995). Keel stirring due to drifting sea ice might generate high frequency internal waves in the pycnocline below the Polar mixed layer and the interaction between tidal motions and steep topography can generate low periodic internal waves in the deep that may initiate interleaving motions (see below).

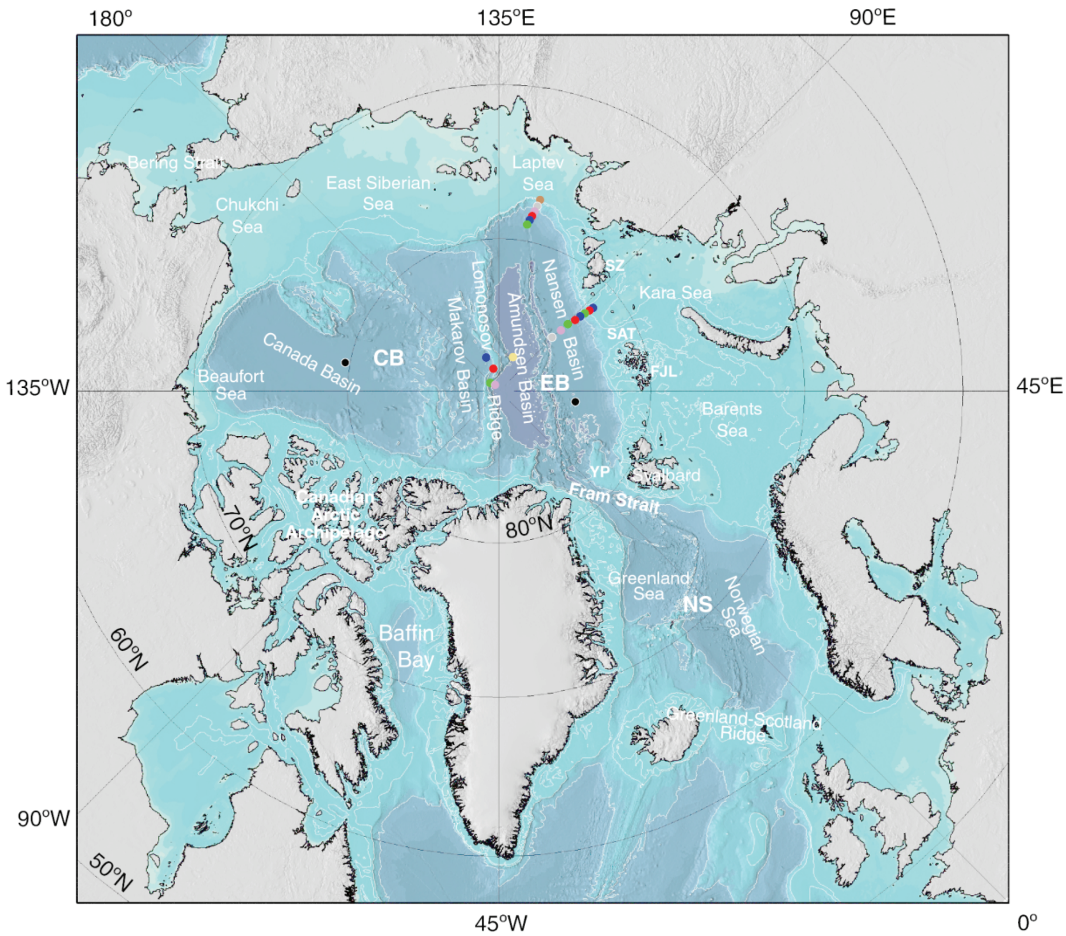

Fig. 1. Bathymetry of the Arctic Mediterranean Sea showing the main features, basins and ridges. CB, Canadian Basin; EB, Eurasian Basin; NS, Nordic Seas; FJL, Franz Josef Land; SAT, St. Anna Trough; SZ, Severnaya Zemlya; YP, Yermak Plaeau. The stations discussed in the text and shown in figures 2 to 6 are indicated on the map. The stations shown in figures 3 and 5 lie on the same section. The bathymetry is based *on Jakobsson et al*., 2008.

Double-diffusive convection and double-diffusively driven intrusions are among the most prominent non-mechanically driven mixing processes. For the present purpose we assume that sea water is incompressible and that its density is determined by two components; its temperature (T) and its salinity (S). The stability due to the vertical distribution of one component e.g. salinity, less saline water above more saline, might be reduced by the temperature stratification, colder water above warmer. Double-diffusive convection is an internal process that releases the potential energy stored in the unstably stratified component, heat or salt, and uses the released energy for mixing and transports. The final configuration will therefore always have lower potential energy than the initial state. The mechanism that releases the energy is the difference in molecular diffusion rates between heat and salt; the coefficient of heat conduction (κ_T) is 100 times larger than the diffusion coefficient of salt (κ_S). This is why a low energy environment is needed. Turbulent mixing, when present, tends to mix heat and salt at the same rate.

Double-diffusive convection is essentially a vertical process and has two distinct modes, the saltfinger mode and the diffusive mode. The existence of saltfingers was first realised, when the problem of pumping colder, less saline and denser water through a tube from the deep ocean to the surface was addressed (*Stommel et al*., 1956). The tube would conduct heat from the surroundings as the water rises in the tube but would prevent salt from entering. This makes the water in the tube warmer and less dense and it continues to rise, once it had been displaced far enough for the ambient stratification to create a temperature difference between the water in the tube and the surrounding water. This also works in the opposite direction. Warmer water displaced downward would be cooled and continue to sink (*Stommel et al*., 1956). When it became clear that the slower molecular diffusion of salt compared to that of heat could fulfill the role of the tube, the theory of saltfinger convection could be formulated (*Stern*, 1960; *Turner*, 1973).

In an ocean stably stratified with warm, saline water above colder and fresher water a vertical perturbation will bring warmer water into colder surroundings and vice versa. The diffusion of heat will cool the warmer water, and heat the neighbouring colder water parcels. The warmer water will become denser and sink, while the colder water becomes lighter and rises. Saltfingers are created. They have vertical scale of one to a few cm and are closely packed, exchanging heat between each other as they rise and sink. The vertical motion does not continue indefinitely, eventually the fingers become unstable and buoyant water convects upward and downward and mixes and homogenises the water above and below the saltfinger interface (*Stern and Turner*, 1969). The vertical scale of the interfaces is usually less than one metre and the interface tends to be thicker if the stability is larger (*Stern*, 1975; *Kunze*, 1987). Thermohaline staircases are thus formed with thin interfaces and strong gradients in temperature and salinity between thicker, almost homogenous layer stirred by convecting parcels. The saltfingers exchange mass between the layers and transport salt more efficiently than heat. The flux ratio $F_{\alpha T}/F_{\beta S}$, α being the coefficient of heat expansion and β the coefficient of salt contraction, relating the fluxes of density due to heat to that due to salt, is ~0.7 (*Kunze*, 1987).

The other mode is diffusive convection, occurring when colder, fresher water is located above warmer, saline water. If a perturbation forces the isotherm (and isohalines) closer together, the molecular diffusion of heat will increase and the water above the stronger gradient will become warmer, the water below colder. The parcels will convect, the warmer upward and the colder downward. They then enter colder and warmer water respectively and become cooled, or heated, and the motion will be stopped and reversed. An oscillatory instability is created (*Veronis*, 1965). The oscillations homogenise the water above and below, creating layers separated by sharp interfaces with strong temperature and salinity gradients. The water immediately above the interface will be heated, become lighter and convect upward, while the water below the interface is cooled and made denser and convect downward. The slower diffusion of salt allows only little salt to be transported from the interface by the convection and the flux ratio at the diffusive interface is $F_{\beta S}/F_{\alpha T} \sim 0.1$, much smaller than at the finger interface. However, no mass is exchanged through the diffusive interfaces, as is the case at the saltfinger interface. The diffusive interfaces thus essentially only allow for molecular diffusion of heat. The ensuing convection, however, prevents the diffusive interface from growing in thickness and the diffusive transports can remain high (*Stern*, 1975).

The transports through the saltfinger and diffusive interfaces are commonly estimated assuming that the transports depend upon the unstable (driving) property step as $(\alpha\Delta T)^{4/3}$ and $(\beta\Delta S)^{4/3}$ respectively. This form ensures that the transports are independent upon the depth of the homogenous layers. The fluxes also depend upon the stability ratios $\alpha\Delta T/\beta\Delta S$ (finger interface) and $\beta\Delta S/\alpha\Delta T$ (diffusive interface) respectively. The ratios are formed with the stabilising component in the nominator, the unstable in the denominator. The fluxes, especially at the diffusive interface, decrease significantly when the stability ratios become large (*Turner*, 1965; *Turner*, 1973; *Schmitt*, 1981).

2. *Diffusive interfaces in the Arctic Ocean*

In the upper 300m of the Arctic Ocean cold, low salinity water is lying above warmer and saline water (Fig. 2), a stratification promoting the formation of diffusive interfaces and double-diffusive convection. The Arctic Ocean was in fact one of the first regions, where diffusive interfaces were observed (*Neal et al.*, 1969). The layers and interfaces were observed from the drifting sea ice in the Canada Basin, a stable platform with minimal wave motion, making the observations possible. The steps were located in the thermocline between the cold Arctic Ocean upper waters and the Atlantic layer. The temperature steps are small and the depths of the homogenous layers range between a few to 10 metres. Closer to Fram Strait, in the Nansen Basin, where the Atlantic water is warmer and the temperature contrast between the Atlantic layer and the overlying water is larger and the homogenous layers are thicker. The stability of the interface is also weaker since the salinity step is smaller. The diffusive interfaces can contribute significantly to the vertical heat flux where, as in the Nansen Basin, the layers thicknesses are large and the temperature contrasts strong (Fig. 2). Since the molecular diffusive flux through the interface is the controlling factor, several interfaces with smaller temperature steps drastically reduce the heat transfer and the vertical heat flux from the Atlantic layer in most part of the Arctic Ocean, and especially in the Canada Basin, is expected to be small.

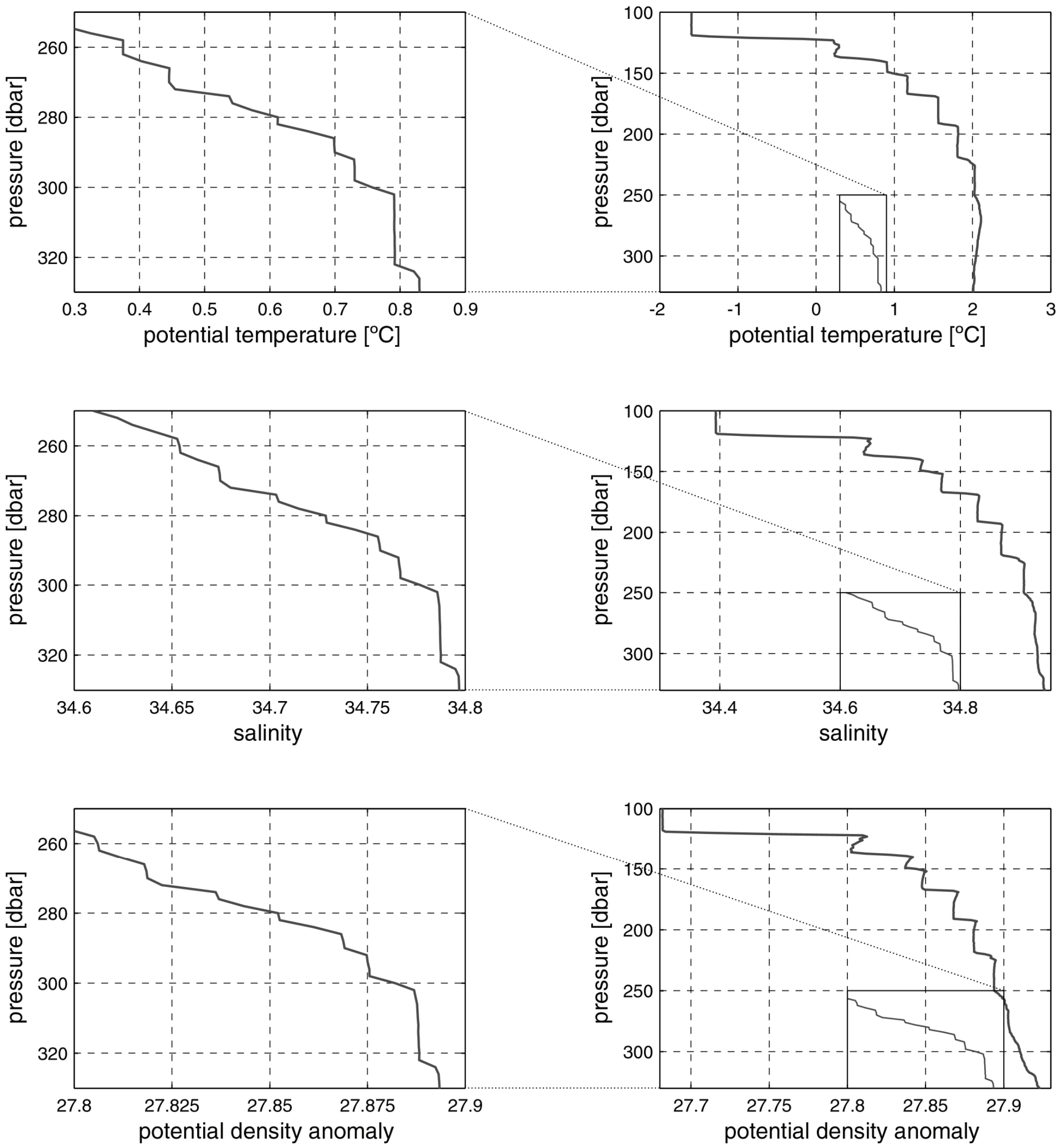

Fig. 2. Potential temperature, salinity and density profiles showing steps. The left panel is from the Canada Basin (data from IB Oden taken on the Healy-Oden TransArctic Expedition (HOTRAX) 2005) and the right panel is from the Nansen Basin (Data from the IB Oden cruise 2001). Notice the differences in the scales.

The mechanisms determining the layer thicknesses have not yet been identified, nor has the process creating the homogenous layers in an environment as stably stratified as the Canada Basin been found. If the initial profiles are linear in both temperature and salinity, and thus also in density, the rise of a buoyant parcel will be restricted by the stratification. Only if the stability ratio $\beta\Delta S/\alpha\Delta T$ is less than 1.5 would the parcels rise (sink) freely and only be limited by the rate of heat diffusion out of the parcels (*Rudels*, 1991). For larger stability ratios the density anomalies created at the interfaces are not capable of rising and homogenise the density gradient into the observed layers. Either the initial conditions must have been different with weaker stability, or the layers are created by a combination of mechanical mixing, e.g. through an unstable shear flow and double-diffusive convection, and then maintained by the diffusive convection.

Unless the reservoir of heat below is unlimited and a continuous cooling is applied at the upper boundary, the temperature steps will gradually diminish. The salinity steps, however, remain more or less unaffected because the small transport of salt, and the stability at the interfaces increases. This lowers the fluxes and could be one reason why the observed TS properties are not useful in finding the initial formation process. Furthermore, as the step becomes smaller the temperature anomaly of the convecting parcels decreases. They may no longer be able to rise through the entire layer before the anomaly has diffused away and the convection will be confined close to the interfaces. The layers are then not thoroughly mixed, and it is possible that a new interface is created in the middle of the layers, increasing the number of interfaces and layers and thus lowering the transports. This might be one reason for the numerous but thin layers observed in the thermocline in the Canada Basin as compared with the thickness of the layers in the Nansen Basin (Fig. 2).

3. The stratification in the intermediate and deep layers of the Arctic Ocean

Below the temperature maximum of the Atlantic layer the temperature decreases to a minimum between 600 and 1200m above the bottom (between 2400–3500m depth) in all basins except the Makarov Basin. The temperature then increases slightly before the homogenous bottom layer is reached. Also here temperature and salinity steps are observed (*Timmermanns et al*., 2003).

The salinity increases towards the bottom in all basins except the Nansen Basin and part of the Amundsen Basin, where a salinity maximum in the Atlantic layer is observed. The salinity maximum is located deeper than the temperature maximum. This is a natural consequence of the stratification shifting from salinity stabilising/ temperature destabilising to temperature stabilising/salinity destabilising. The change in temperature stratification has to occur first to avoid overturning. In similar manner a change from temperature stabilising/salinity destabilising to a salinity stabilising/ temperature destabilising stratification would have the salinity minimum above the temperature minimum.

The unstable salinity gradient has two causes, the high salinity of the Atlantic water entering through Fram Strait and the inflow of colder, less saline water along the St. Anna Trough. This is part of the second inflow from the Nordic Seas to the Arctic Ocean, The Barents Sea branch. The water becomes cooled and freshened as it passes over the Barents Sea and it penetrates into the Arctic Ocean in the depth range between 200 and 1200m, occasionally also deeper (*Schauer et al*., 1997, see also Fig. 3). This implies that there is a region, where both components are stably stratified, and then below the salinity maximum another interval where the salinity is destabilising, and the possibility for salt fingers convection and formation of staircases exists. Below about 1000m the salinity profiles are again stably stratified.

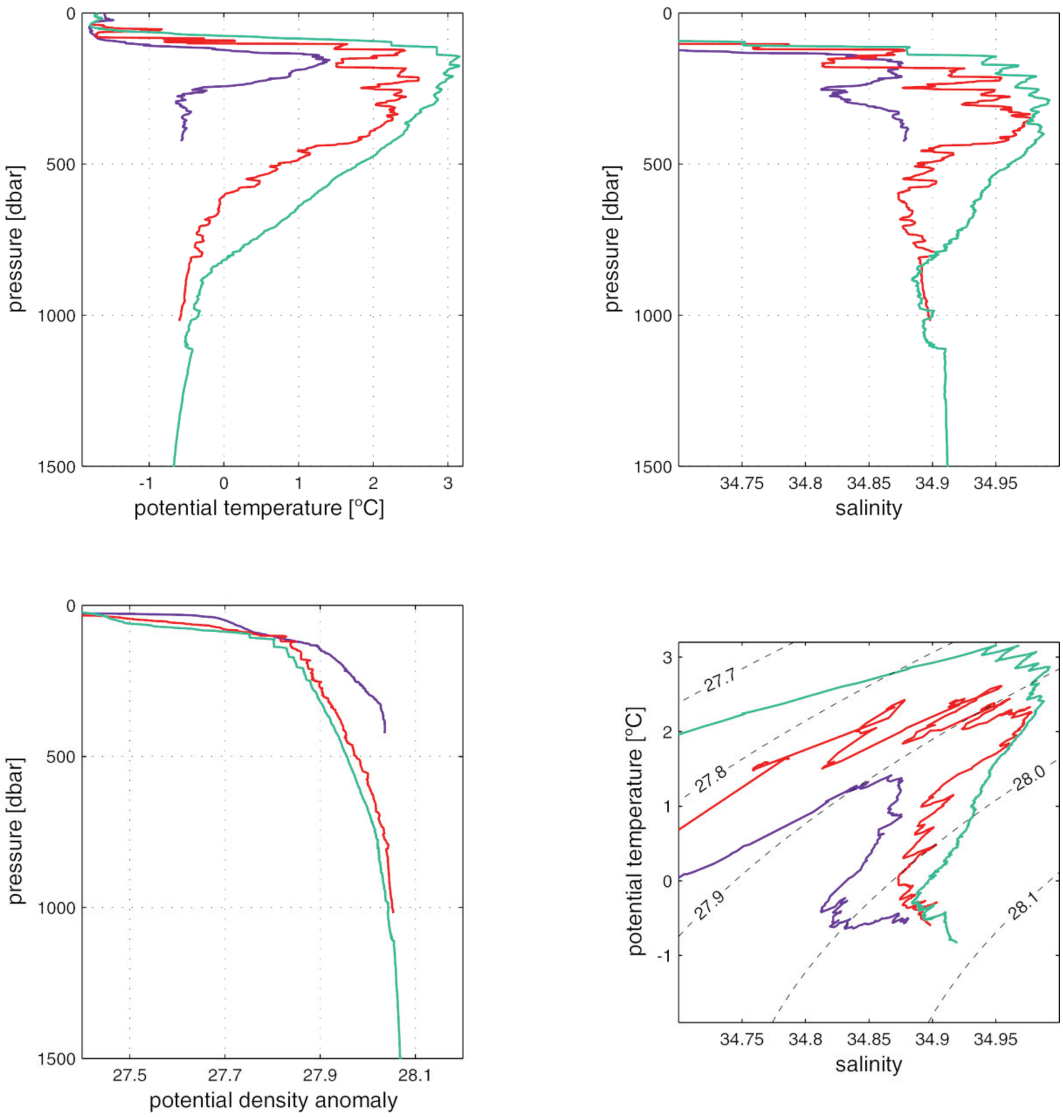

Fig. 3. Potential temperature, salinity and potential density profiles and θS curves from the eastern Kara Sea slope (Data from the Synoptic Pan-Arctic Climate and Environment Study (SPACE)/IPY cruise with RV Polarstern 2007). Notice the denser Barents Sea branch water.

4. *Thermohaline intrusions in the Arctic Ocean*

Thermohaline staircases are, however, not what dominates the Atlantic and the intermediate layers of the Arctic Ocean, but rather the existence of inversions in both salinity and temperature and the presence of intrusions and interleaving layers. The alternating maxima and minima in the vertical indicate interleaving between two different water masses, one warm and saline, the other cool and fresh. Between the layers, and the two water masses, alternating diffusive and saltfinger interfaces are formed. A warm intrusion has a saltfinger interface below and a diffusive interface above. For a cold intrusion the situation is reversed. Both interfaces transport heat and salt at different rates and the density of the intrusions changes. This leads to buoyancy forces that can enforce motions across the front from one water mass into the other and this, double-diffusively driven, motion can enhance the lateral mixing between the two water masses. Depending upon the property steps and the stability ratios at the interfaces either the transfer of density through saltfinger interfaces or through diffusive interfaces may dominate. If the saltfinger fluxes dominate, warm and saline intrusions will rise and cold and fresh intrusions will sink relative to the isopycnals as they move across the front. When the heat transported through the diffusive interfaces is most important warm, saline intrusions will sink while cold, fresh intrusions will rise.

The existence of thermohaline intrusions in the Arctic Ocean was first reported by *Perkin and Lewis* (1984) north of the Yermak Plateau and then by *Quadfasel et al*. (1993) and *Rudels et al*. (1994). Their interpretations of the formation of the intrusions were, however, different (see below). The interleaving layers observed in the Arctic Ocean are perhaps the most extensive found in the world ocean (*Carmack et al*., 1997). They are observed in all possible stratifications, saltfinger unstable, diffusively unstable and also when both components are stably stratified. If the intrusions have a dynamics of their own, they could contribute significantly to the spreading of heat from the boundary current at the slope to the interior of the basins. *Carmack et al.* (1997) considered the possibility that the layers extended over the entire Arctic Ocean and that their contribution in spreading heat into the basins was considerable. In later work (*Walsh and Carmack*, 2002; *Walsh and Carmack*, 2003) the estimate of the lateral spreading of the heat was reduced but still considered important. *Walsh and Carmack* (2003) also estimated the advection in the intrusive layers to ~0.001 ms^{-1}.

In the Arctic Ocean observations indicate that in general cold, fresh intrusions rise and warm, saline intrusions sink relative to the isopycnals, indicating the dominance of the diffusive fluxes. However, often the interleaving layers are parallel to the isopycnals, suggesting little or no cross frontal motion. The intrusive motions are expected to cease, when the driving steps and the transports become small, or perhaps earlier, if a balance between the fluxes through the two interfaces can be achieved.

That intrusive layers can form in a frontal area with density compensating horizontal gradients has been shown by linear stability analysis (*Stern*, 1967; *Toole and Georgi*, 1981, *McDougall*, 1985a; *McDougall*, 1985b). To get an infinitesimal disturbance to grow one component has to be unstably stratified, and the transports of the stabilising component has to be smaller than the flux of the unstably stratified component. In a basic saltfinger stratification an infinitesimal disturbance towards colder, less saline and less dense surroundings will enhance the saltfinger fluxes below and weaken them above the disturbance. The water parcel becomes less dense and the disturbance will grow and continue across the front. The colder and fresher water immediately above and below will become denser and move in the opposite direction, towards warmer, more saline and denser surroundings. Interleaving layers are formed. As the intrusions penetrate farther diffusive interfaces develop above the warm saline intrusion (*McDougall*, 1985a, b). In a basic diffusive stratification the situation is the opposite and warm, saline intrusions would become cold, less saline and denser and sink. In the Arctic Ocean the diffusive buoyancy fluxes appear to dominate and warm, saline intrusions sink through the isopycnals.

May and Kelley (2001) adopted this approach to explain the formation of the intrusions reported by *Perkin and Lewis* (1984). They assumed an interaction between Atlantic water just entering the Arctic Ocean through Fram Strait and old “Arctic” Atlantic water in the interior of the basin. They also added the possibility that vertical shear of the along front background velocity, the baroclinicity, also contributed to the formation of the intrusions (*Kuzmina and Rodionov*, 1992; *May and Kelley*, 1997).

The stability analysis does not work when both components are stably stratified and no energy is initially available to be released and generate the intrusions. In this case the frontal zone has to be narrow to allow finite disturbances to penetrate into the opposite water mass and create the initial inversions needed to generate vertical double-diffusive transports. Such systems have been studied in the laboratory and intrusive layers have been observed (*Ruddick and Turner*, 1979; *Holyer et al.*, 1987). For the Arctic Ocean another process; differential diffusion, the fact that heat diffuses more rapidly than salt, has also been suggested as a possibility to create intrusive layers without initially releasing any stored potential energy (*Merryfield*, 2002). The problem of creating thermohaline interleaving at a finite frontal zone with different background stratifications is presently addressed in another work and will not be considered further here (*Rudels, Kuzmina, Stipa and Zhurbas*, in prep).

We concentrate on the situation in the Nansen Basin, where the two Atlantic inflows, through Fram Strait and over the Barents Sea, meet north of the Kara Sea and flow parallel in the boundary current along the Eurasian continental slope. The large property contrasts between the two inflows create a strong front between the two branches, similar to the system studied by *Ruddick and Turner* (1979). This eventually leads to rather chaotic interactions between the two branches, especially in the Atlantic core, and strong inversions are created, in particular north of Severnaya Zemlya (*Rudels et al.*, 1999; *Rudels et al.*, 2000 (see also Fig. 3)). The interleaving gradually loses its chaotic character and becomes more organised farther downstream, north of the Laptev Sea (Fig. 4). More regular intrusions are observed in the thermocline and above the salinity maximum in the Fram Strait branch farther offshore, where the Barents Sea branch has not penetrated into the warm Atlantic core (Fig. 3). This could indicate that the shallower part of the Barents Sea branch interacts more rapidly with the Fram Strait branch than do the Atlantic and intermediate layers.

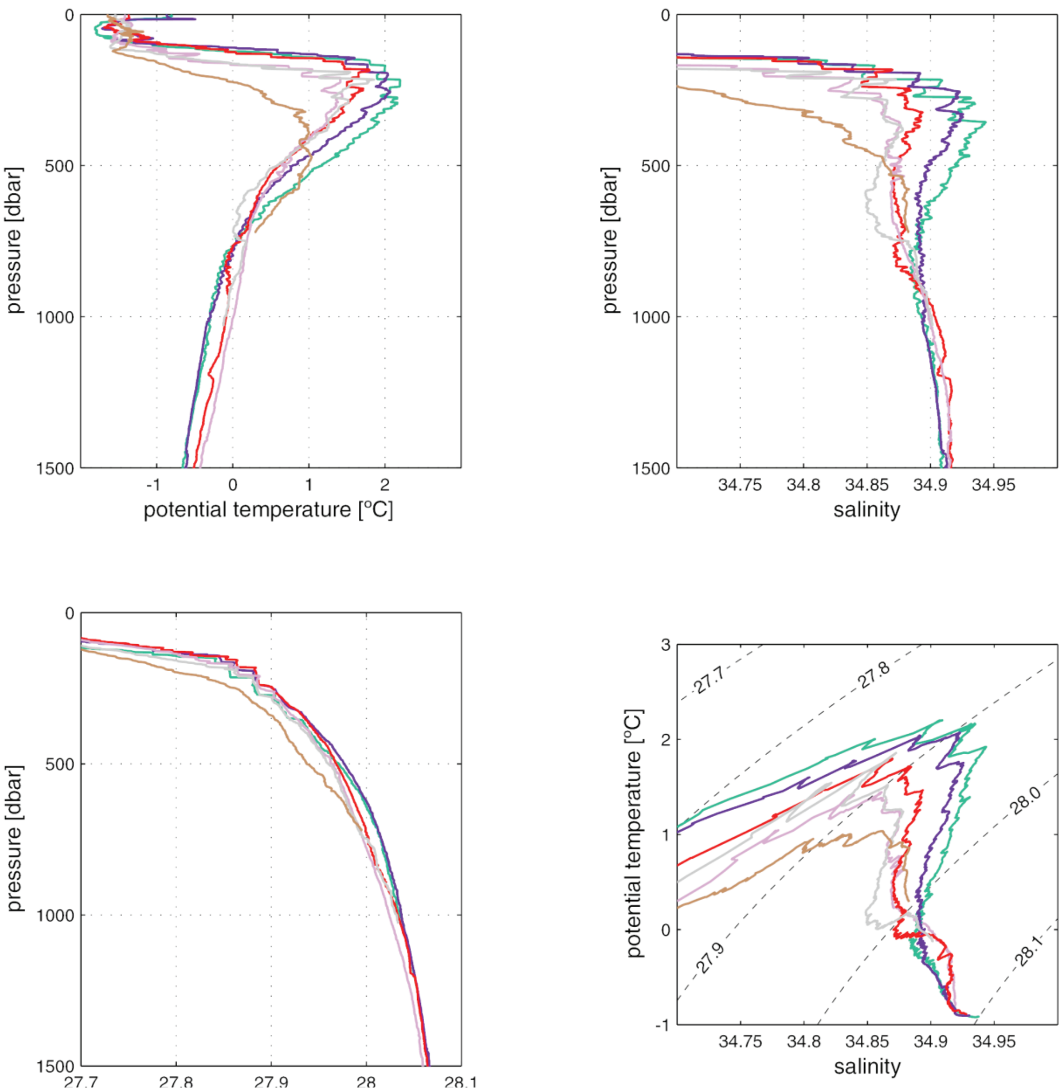

Fig. 4. Potential temperature, salinity and potential density profiles and θS curves from the Laptev Sea slope. (Data from the SPACE/IPY cruise with RV Polarstern 2007). Notice also the step structure in the density profiles, the diffusive interfaces being more stable.

On the basin side of the Fram Strait branch core more organised intrusions are observed, in the warm Atlantic core and in the intermediate waters below. The structures found here are similar to those seen north of the Laptev Sea. It is known that the Barents Sea branch water dominates in the Amundsen Basin, implying that the Barents Sea branch is partly deflected from the slope towards the interior of the basin north of the Laptev Sea (*Rudels et al*., 1994). It is then likely that the Fram Strait branch water, being located on the basin side of the Barents Sea branch, also takes part in this deflection and becomes confined between the Barents Sea branch water at the slope and in the Amundsen Basin. If the intrusions observed between the two branches at the slope are not removed by diffusion, which is unlikely considering the short distance, they will be advected with the front between the branches and be found in the Nansen

Basin on the basin side of the Fram Strait branch, moving in the opposite direction. This is the interpretation made by *Quadfasel et al*. (1993) and *Rudels et al*. (1994).

In the Nansen Basin fronts in the properties are clearly seen between the Fram Strait branch and the Barents Sea branch water on the basin side of the Fram Strait branch, and cross frontal transports are likely to occur (Fig. 5). However, in the Amundsen Basin the frontal structure is weak and the profiles and the TS curves practically lie on top of each other (Fig. 5, see also e.g. *Rudels et al*., 1999). This implies little cross frontal transfer and the intrusions are advected with the main circulation gyre and do not contribute to the spreading of heat into the interior of the Arctic Ocean basins by their own dynamics.

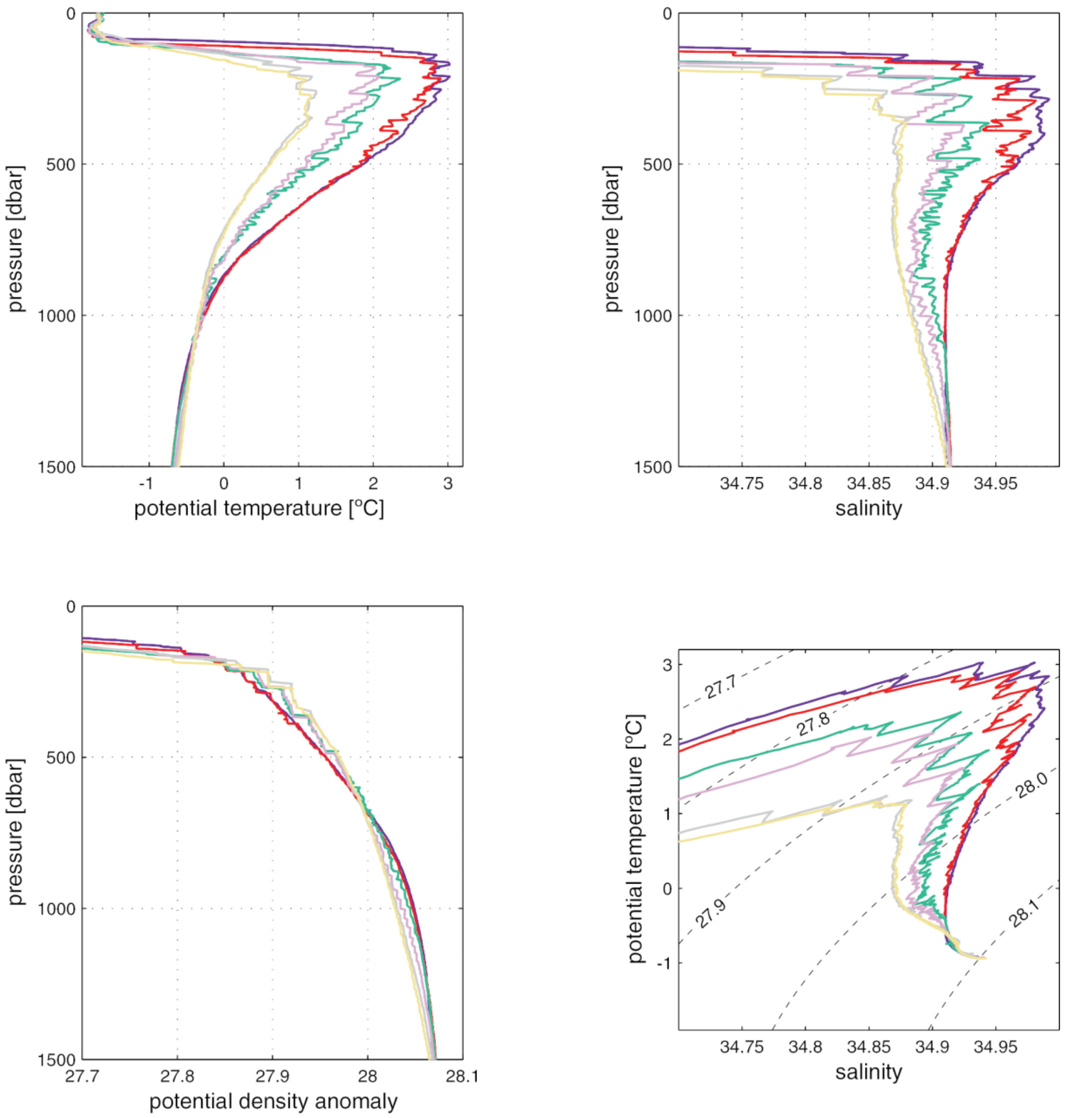

Fig. 5. Potential temperature, salinity and potential density profiles and θS curves from the Nansen and Amundsen basins. (Data from the SPACE/IPY cruise with RV Polarstern 2007). Notice also the step structure in the density profiles, the diffusive interfaces being more stable.

That intrusions develop quickly can be seen from eddies formed at intermediate depth in the Nansen Basin as the two branches meet. Another striking example is the penetration of Makarov Basin deep water into the Amundsen Basin at 2000m depth through an intra-basin in the Lomonosov Ridge (*Björk et al*., 2007, see also Fig. 5). Both the Makarov Basin and the Amundsen Basin are in this depth range stably stratified in both temperature and salinity. Nevertheless intrusions are formed almost directly as the deep water from the Makarov Basin enters the Amundsen Basin and gets into contact with the Amundsen Basin deep water. The Makarov Basin deep water and the intrusions do not penetrate directly into the Amundsen Basin but remain at the Lomonosov Ridge slope and become advected towards Greenland (*Björk et al*., 2007).

The implications are then that the interleaving layers are created primarily at frontal zones and the stronger the front, the more intense the interleaving. Double-diffusive convection might not be the process initially setting the scale of the layer thickness and create the first inversions. In the case of both component being stably stratified finite lateral disturbances are necessary. These might arise from more dynamic processes like baroclinic instability or internal waves and internal tides. However, double-diffusive convection is likely to contribute to the propagation of the intrusions into the opposite water mass. Double-diffusive convection might also remove some of the initially formed interfaces by causing thin layers to merge (*Ruddick*, 1984). The transport of buoyancy through one interface i.e. the diffusive interface, could at small stability ratios so strong that the stability at the neighbouring finger interface decreases and disappears, causing the layers to overturn and merge, and thus create the more “organised” structure of the intrusions seen in the interior of the basins. The step like structure seen on some of the saltfinger interfaces might be evidence of such process (see e.g. Fig. 5).

When the double diffusive transports diminish as the gradients become smaller, the motion in the layers would eventually stop. The inversions might, however, still remain, at least in one component, and would be convected with the mean gyre circulation. This could be one reason for the extensive spreading of inversions and the intrusive layers. The circulation in the Arctic Ocean consists of several loops and gyres largely steered by the bathymetry. The intrusive layers appear to align across these loops and extend between the different basins despite the different time scales of the circulation in the individual loops. Is this just an illusion? The interleaving structures are created at the depth and density intervals where the contrasts between different water masses are the largest. This is a fairly limited region in TS space and the intrusions will be located in similar density intervals even if they are created at different times and in different parts of the Arctic Ocean. The fact that many intrusions from different parts of the Arctic Ocean are aligned then does not necessarily mean that they are part of the same intrusive feature. In that case the importance of the thermohaline intrusions in spreading heat into the interior of the ocean basins might be smaller than anticipated and considerably smaller than the advective transports in the main gyres. However, to describe and understand the physics and the formation of the double-diffusive intrusions in the Arctic Ocean still remains an important and intriguing problem.

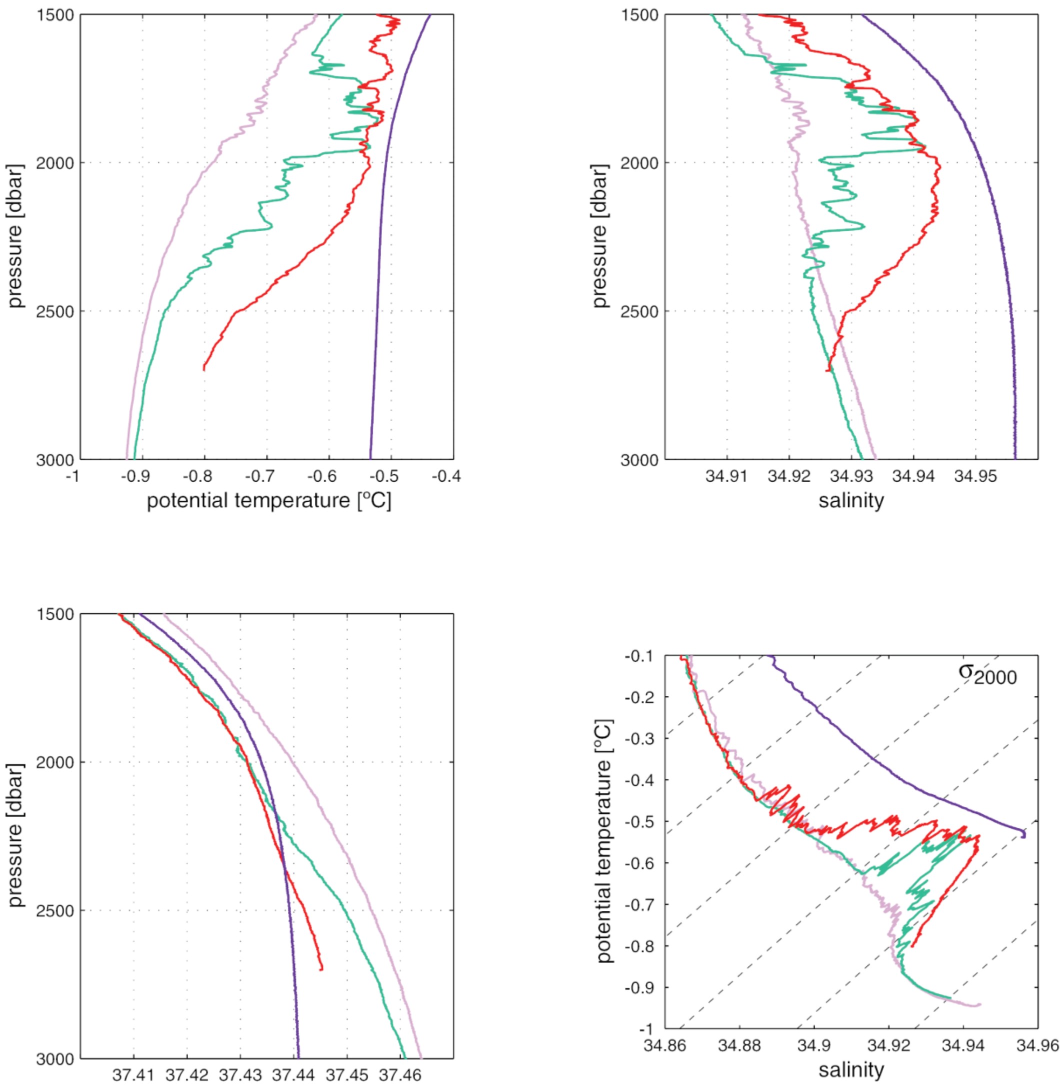

Fig. 6. Potential temperature, salinity and profiles and θS curves from the Makarov Basin (blue), the Intra-basin in the Lomonosov Ridge (red), the Amundsen Basin side of the Lomonosov Ridge (green) and the Amundsen Basin (Magenta). (Data from IB Oden taken on the HOTRAX cruise 2005).

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