Formation and export of water masses produced in Arctic shelf polynyas - process studies of oceanic convection

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The contribution of brine-enriched bottom water from Arctic shelves to intermediate and deep water masses of the adjacent Arctic Ocean or the Nordic Seas is a widely discussed topic in Arctic Oceanography. This paper presents an overview of process-oriented modelling which was conducted to deepen our understanding of oceanic convection and its role in water mass formation. It arrives at a conceptual picture of the convective formation of bottom water masses in Arctic shelf seas as a consequence of ice-ocean interactions and considers the role of sediments in slope convection. To investigate and discuss the processes which take part in transformation, production and export of dense shelf water masses a hierarchy of numerical models of different type and spatial resolution was applied. The models cover spatial scales from well below the internal Rossby radius of deformation up to the mesoscale.

Key words: Arctic Sea, water formation, water export, polynyas, convection.

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Introduction and overview

Intermediate and deep waters of the World Ocean are generally separated from direct contact with the atmosphere by a stable perennial thermo- or halocline. There are only a few exceptions where convection in the open water column driven by atmospheric forcing may penetrate to intermediate depths or even to the bottom of an ocean (Killworth, 1983). In the northern hemisphere open ocean convection occurs in the Labrador and Greenland Seas (Carmack and Aagaard, 1973; Clarke and Gascard, 1983), in mid-latitudes in the Mediterranean Sea (Carmack and Aagaard, 1983), in mid-latitudes in the Mediterranean Sea and in the southern hemisphere in the Weddell Sea (Rudels, 1993; Gordon, 1978). Only deep convection regions in the north contribute to and maintain source waters for the global “Conveyor Belt” circulation (Broecker, 1991). In the Arctic Ocean both a perennial halocline and ice cover prevent deep-reaching convection in the open water column (Aagaard et al., 1985). Slope convection, elsewhere also called “cascading” (Blindheim, 1989; Huthnance, 1995), of dense bottom water masses produced on the shelves around the Arctic Basin provides a substantial contribution to renewal of intermediate and deep water masses beneath the halocline (Blindheim, 1989; Schauer, 1995).

Dense shelf bottom water (Midttun, 1985; Aardal and Loeng, 1991) is formed mainly by locally confined thermo-haline convection events that follow an erosion of summer stratification on the shelf. Openings in the ice cover, i.e. “polynyas”, allow intense thermodynamic ice formation and the necessary brine-release for the onset of haline convection (Rudels, 1990). The brine-enriched bottom water forms a bottom-arrested gravity plume which follows topographic depressions. Finally, the plume exits the shelf on the continental slope and enters deep ocean basins at its horizon of equilibrium density.

The importance of the dense waters produced on Arctic shelves has already been inferred from analyses of water masses and tracers of the Arctic Ocean (Aagaard et al., 1981; Swift et al., 1983; Aagaard et al., 1985; Rudels, 1986; Anderson et al., 1985; Bauch et al., 1995). However, the integral contribution of the shelves to water mass formation in the Arctic Ocean cannot be estimated from the few available direct observations on production and slope convection of shelf waters (Quadfasel et al., 1988; Schauer, 1995; Weingartner...
et al., 1996; McDonald et al., 1996). Attempts to quantify the contribution of shelves by integral parameters derived from remotely sensed space–time ice statistics of polar shelf polynyas (Martin and Cavalieri, 1989; Cavalieri and Martin, 1994), may suffer from an insufficient knowledge of the temperature and salinity (TS) properties of shelf waters (Backhaus, 1995; Backhaus et al., 1996) as will be demonstrated below. Recent model simulations, supported by geological evidence (Fohrmann, 1996), suggest that sediments, suspended in bottom-arrested gravity plumes, may play a noteworthy role in water mass formation caused by slope convection.

A combination of the model simulations described below with available observations and our present knowledge about Arctic processes suggests four phases in the formation and export of shelf water masses (Backhaus, 1996; Backhaus et al., 1996) in which both convection within the water column and slope convection play an important role. Before looking at these processes in more detail below a cartoon (Fig. 1) provides a schematic overview. The following sections (phases 1–4) will refer to this cartoon.

Phase 1: large scale pre-conditioning

In early winter a layer of fresh surface water caused by river run-off or ice melt during summer inhibits deep reaching (i.e. penetrating) convection on the Arctic shelves. The results of simulations with a general circulation model (GCM) for the Barents Sea (Harms, 1997) suggest that substantial production of dense bottom waters will only occur in late winter because it takes several months, about two-thirds of a winter, to erode convectively the pronounced seasonal halocline by cooling and release of brines from freezing sea-ice. Only if the water column is largely homogenised, or at least weakly stratified, may formation of dense bottom water by penetrating convection occur. In late winter, however, an ice cover has been established which hinders further convective water mass formation. Therefore, particularly in late winter, episodic occurrences of open water appear to play an important role in regard to production of dense shelf bottom water masses. The import of warm water masses of Atlantic origin onto the Barents shelf may cause a negative feedback on both deep reaching convection and ice-growth. This effect will be far less pronounced in Eurasian shelf areas because there a significant import of Atlantic water does not occur.

Phase 2: convective production and small scale pre-conditioning

In the course of episodic breakouts of cold and dry polar air, latent heat shelf polynyas (Smith et al., 1990) form in the lee of topographic obstacles (e.g. islands, headlands) or in the vicinity of a solid, land-fast ice edge (Fig. 1a,b). These “lee-polynyas” appear in satellite images, conventional ice charts and model simulations (Harms, 1977). The atmospheric episodes are concomitant with extreme fluxes of momentum, heat and moisture due to low temperature, typically well below −20°C, low humidity of the air, and high wind speeds. The oceanic (latent plus sensible) heat-loss may be in the order of 1000 W m⁻² (Moore, 1996). During such an event thermo-haline convection, driven by both heat loss and brine-release from freezing sea-ice, may account for a total homogenisation of the initially weakly stratified shelf water column. It is assumed that these synoptic events, which typically last for 3–5 days, may account for an effective convective formation of shelf water masses (Martin and Cavalieri, 1989; Cavalieri and Martin, 1994). Convection that did not penetrate through the entire water column may pre-condition the water column for a subsequent convective episode (Fig. 1b). Waters under the closed ice cover around a polynya remain stratified because they are not exposed to the atmosphere. There, due to the insulating effect of the ice, convection will be much weaker or may not exist at all.

Phase 3: re-stratification

With the end of a cold air breakout the ice closes over the polynya. Polynya waters form a homogenised water body that is denser than the surrounding weakly stratified shelf waters. This body may cover the entire water column, or parts of it, depending on the final penetration depth of convection. The convection region will have a similar spatial scale as the polynya, which is generally larger than the local internal Rossby radius of deformation that for polar shelves is typically a few kilometres. Waters under a closed polynya will re-stratify because the dense water body collapses under the influence of both gravity and rotation. Re-stratification will either form a bottom plume (Fig. 1c) or an internal (anti-cyclonic) eddy (Fig. 1d), depending on penetration depth of convection. Motion of a plume of dense bottom water on an inclined shelf will be retarded by both rotation and bottom friction. Baroclinic instabilities may cause a lateral transport of heat and salt for both plume and internal eddy (Gawarkiewicz and Chapman, 1995). Successive breakouts of polar air with high winds from preferred directions (e.g. in the Barents Sea primarily northerly winds) would eventually result in an effective but localised production of dense shelf waters. This will largely be confined to and governed by the occurrence, position and scale of a polynya. A weak or missing background advection would favour both pre-conditioning and convective water mass production.
Figure 1. Cartoon of processes involved in the formation and export of shelf water masses. (a) Convection covers entire water column within an open polynya, surrounding waters remain unchanged; (b) same as (a) but convective penetration is less than water depth; (c) gravitational rotational collapse of dense polynya water causes re-stratification under closed polynya; (d) collapsing polynya water forms an anti-cyclonic internal eddy whilst waters re-stratify; (e) gravity plume of dense shelf bottom water approaches continental slope; (f) injection of sediment laden gravity plume on continental slope into equilibrium level at intermediate depth; settling of sediments causes upward directed internal convection.
Phase 4: export

Once a slowly moving plume of dense bottom water (Fig. 1e) has reached the edge of the continental slope it will cascade down the slope to an equilibrium depth which depends on both its initial water mass properties and the entrainment of ambient water masses during descent. In contrast to the slow motion on a gently sloping shelf, a plume becomes more energetic on a steep continental slope due to gravitational acceleration. This causes increased entrainment and the plume acts as a “transport vehicle” for the properties of ocean waters entrained during its descent. For instance, slope convection initiated by cold and saline shelf bottom plumes, may entrain properties from the warm and saline Atlantic inflow to the Arctic Ocean which causes a net downward transport of both heat and salt. Penetration of cold and saline bottom-arrested plumes into warmer oceanic waters may be influenced by the thermobaric effect. Sediments initially suspended in a gravity plume or due to erosion on its descent may account for a larger negative buoyancy in comparison to classical water mass plumes. Their higher dynamics cause a different entrainment of water masses and allow for a deeper penetration. Once a sediment plume has reached its equilibrium level reduced dynamics allow for a settling of sediments (Fig. 1f). The plume may then become lighter than the ambient oceanic water masses. This would initiate upward directed convection in the water column. Localised areas of high sediment accumulation at the lower continental slope and intermediate and bottom nepheloid (high attenuation) layers provide evidence of sediment plumes (Blaume, 1992; Rumohr, 1996). Nepheloid layers were also observed in the vicinity of Arctic continental slopes (U. Schauer, pers. comm., 1995).

The intention of this paper is to describe these Arctic processes by presenting results of process-oriented model simulations for the phases 2 to 4 (Fig. 1) which were conducted to gain more insight into the intricate nature of convection and ice-ocean interactions in polar shelves. Convective water mass formation, ice production and brine-release in a hypothetical wind-induced latent heat lee-polynya in an Arctic shelf were investigated by means of a coupled ice-ocean convection model (IOCM). The IOCM resolves spatial scales well below the Rossby radius of deformation. The gravitational collapse of convectively formed polynya water and the export of shelf waters by slope convection, which may be enhanced by suspended sediments, was studied by means of reduced-gravity plume models. The latter resolve scales in the order of the Rossby radius of deformation.

The coupled ice-ocean convection model (IOCM)

In this process study we hypothesise that production of dense shelf bottom waters depends primarily on episodic atmospheric events and on processes with rather small temporal and spatial scales. Therefore, we place our main emphasis on processes within a lee-polynya. The governing equations of the model and technical details regarding its numerical scheme are only outlined briefly below because the main intention of this paper is a description and discussion of the physical processes that emerge from the model results.

The non-hydrostatic ocean model (Backhaus, 1995; Kämpf, 1996) is based on the non-linear, primitive Boussinesq equations for an incompressible fluid and predicts perturbations from a hydrostatic background. Two versions of the model were applied: a computationally economical 2.5-dimensional (SLICE) and a fully three-dimensional version (CUBE) with a high computational demand. SLICE works in a vertical XZ-plane which extends until infinity normal to its plane, the Y-axis. In both model versions Coriolis acceleration is incorporated in the equations of motion for all three space coordinates. The SLICE model, however, is not able to simulate a geostrophic adjustment between flow and pressure field due to missing resolution normal to the model plane where all gradients vanish. Prognostic equations for salinity and potential temperature together with a non-linear equation of state and a non-linear equation for the freezing temperature as function of salinity are applied for the prediction of the density field. The model includes the “thermobaric effect” due to the compressibility of sea water which is most pronounced at low temperatures. A adjustment between non-hydrostatic pressure and flow field requires a simultaneous solution of a Poisson equation for the pressure at each time step. The prescription of lateral boundary conditions is avoided by an application of cyclic boundary conditions, i.e. disturbances that leave the model domain through one boundary and re-enter the domain through the opposite boundary.

Sub-grid scale turbulence is parameterised by means of a simple diagnostic turbulence closure scheme proposed by Kochergin (1987). It assumes that the rate of turbulence can be described by an instantaneous balance or imbalance between shear production derived from the predicted velocity field and stability expressed by the buoyancy frequency, $N^2$. A turbulent Prandtl number of unity is assumed following observations under drifting Weddell Sea pack ice (McPhee and Martinson, 1994). Both SLICE and CUBE are defined on an isotropic, i.e. equidistant grid, with a grid size of 10 m, thus avoiding distortions of convective dynamics and of the parameterisation of sub-grid scale turbulence by the choice of a non-uniform grid. Since no preference is given for any of the respective coordinates equal eddy viscosity and diffusivity coefficients are applied for both horizontal and vertical coordinates. The diagnostic closure scheme yielded values in a range from 0.005 to 0.05 m² s⁻¹ for the turbulent eddy viscosity (and diffusivity) when
applied to convective dynamics in our high-resolution model. These are close to values obtained from observations by M. Pfeef and M. Martinson (1994). The hydrodynamic convection model is interactively coupled with a simple ice model. Models communicate via fluxes of momentum, heat, and salt. The coupled IOCM is mainly applied to conditions where formation of thin, young sea-ice yields thicknesses that are on average far less than one metre. In order to be consistent with the resolution applied for convective ocean dynamics, much smaller spatial scales in ice thermo-hydrodynamics (order of deca-metres) than in previously reported ice models (Semtner, 1976; Hibler, 1979; Parkinson and Washington, 1979; Lemke et al., 1990) are considered.

The initial ice-growth in an ice-free ocean surface is determined from the rate of super-cooling via the amount of released latent heat. The same procedure, but with an opposite sign, is applied for the melting of thin ice. Until the ice thickness exceeds a critical value, which (intuitively) was set to 1 cm, the ice is treated as an open ice. Hence, ice beneath the critical initial thickness will move with the surface currents and will not have an insulating effect on the ocean. With the ice thickness exceeding that critical value its further growth is determined from a one-layer heat conduction equation for thermodynamic ice-growth in which a snow cover is ignored. The wind and ocean stresses acting on the ice are determined from a non-linear bulk formula (Gill, 1982) for the momentum flux.

The ice-ocean system in our IOCM is forced by a homogeneous atmosphere by prescribing a constant ambient air temperature, humidity, and wind speed. For the parameterisation of fluxes of latent and sensible heat and momentum at the sea surface standard bulk formulae were used (Gill, 1982). Short-wave radiation was excluded because only winter conditions are considered, but long-wave (black body) radiation was included for the radiation balance at both ice and ocean surface (M. Maykut, 1986).

The prognostic equation for the ice thickness in the IOCM is formulated in conservative flux form. In our first model experiments, however, excessive convergent ice-growth occurred. It was caused by a superposition of convective surface dynamics with the wind-driven Ekman regime. In the vicinity of convective plumes, which imply a high local convergence of surface currents, the ice thickness grew to several metres within only a few hours. The unrealistically high convergent ice-growth was effectively avoided by introducing a parametric ice mechanic. It considers the fact that the degree of freedom of motion of an ice floe is reduced once it “feels” the presence of adjacent floes in a surface convergence. The ice mechanic was defined by interpreting our own observations from the Greenland Sea. Pancake ice drifting in the open ocean far away from a coast never piles up but rather agglomerates in irregular patches or streaks.

Convective water mass formation in a latent heat lee-polynya

This study on convection and ice formation within a hypothetical latent heat lee-polynya on a polar shelf considers processes with very small spatial and time scales which rarely have been observed directly due to harsh winter conditions in the Arctic. Therefore, a number of simplifying and restricting assumptions had to be made:

- the scale of the polynya must be large compared to the model domain (homogeneity);
- effects of lateral boundaries (fast ice, ice edges, coast lines etc.) are ignored;
- advection and mixing by the large scale flow is neglected;
- an initially ice-free ocean surface is assumed; and,
- the ocean does not interact with the atmospheric boundary layer.

For the sake of computational economy most model experiments were conducted with the simpler 2.5-dimensional SLICE version. The CUBE model served to confirm validity of model results produced by SLICE. Two experiments were conducted to study the effect of convective pre-conditioning in the event that subsequent breakouts of polar air would cause a lee-polynya at the same location. It was assumed that the waters, modified by local convection in the polynya, remain at that location.

In the first experiment the IOCM was initialised with a single TS-profile (invoking assumption of homogeneity) for a water column of 250 m depth. The profile was predicted by a GCM (Harms, 1997) for a lee-polynya in late winter in the Barents Sea near Franz Josef Land. The ocean in the IOCM was initially at rest, its surface ice-free and the applied atmospheric forcing was built up over 6 h in order to spin up the model. A breakout of polar air was simulated by prescribing a constant homogeneous wind with a speed of 10 m s⁻¹ and an ambient air temperature of −20°C with 50% saturated air. This forcing was applied for a duration of 160 h. Given the predicted sea surface temperatures the forcing resulted in a total oceanic heat-loss of about 1000 W m⁻² due to a superposition of sensible (−600 W m⁻²) and latent heat fluxes (−300 W m⁻²), and long-wave radiation (−80 W m⁻²).

In order to study the result of an atmospheric event which hits a convectively pre-conditioned water column, a second experiment was conducted with the same atmospheric forcing. The final TS-fields predicted by the first simulation were laterally averaged over the model domain to filter out any remnants of the preceding convection. The resulting mean TS-profile was then used...
to initialise the second IOCM experiment in which, again, an ice-free surface and an ocean at rest were prescribed. Note that this experimental set-up implies a freshwater loss for the polynya region since the ice formed during the first experiment has been removed.

The predicted water mass formation for both experiments can best be described by means of a time series of TS-diagrams (Fig. 2). Distinct water bodies were formed during the evolving convection as shown by the changing TS-properties at certain stages of the process. After 60h of simulation time the predicted TS-properties of the first experiment already deviated considerably from the initial water mass properties (Fig. 2a). Thermal convection prevailed during this period and no, or only little, ice was formed. This is seen from a comparison of the TS-properties at 12h (still close to initial properties) with the TS-cluster obtained after 60h (Fig. 2a). The upper part of the water column, temperatures around $-1.3^\circ$C, became saltier and warmer due to both convective mixing and a net upward heat transport caused by convection. Surface waters, which were initially close to the freezing point (T: $-1.85^\circ$C, S: 34.7), became warmer by ca. 0.2$^\circ$C, thus inhibiting ice formation, which is seen from a temperature scatter around a value of $-1.6^\circ$C (cf. Fig. 2a). Surface salinities increased from 34.7 to about 34.75 because saltier waters from below were mixed upward by convection. Water masses near the seabed, with initial temperatures around a value of $-1.1^\circ$C, i.e. well above the freezing point, and with salinities above 34.8, remain largely unchanged at this stage. However, a slight mixing around the initially discrete data points (cf. Fig. 2a) is evident although penetrative thermal convection did not reach the lower parts of the water column. Convection originating at the surface caused non-hydrostatic pressure fluctuations which are transmitted through the entire water column. The mixing below convective penetration depth was caused by motions which were driven by these pressure fluctuations.

The final stage of the water mass formation of the first experiment is given by the TS-cluster after 156h (cf. Fig. 2a). Now, across much of the TS-diagram salinity has increased further, partly due to brine-release from freezing ice. The near surface waters (lower left end of the point-cluster) have been cooled down to about $-1.8^\circ$C and a homo-haline, finger-like T-S-cluster approaches the freezing point. Cold melting water at the surface caused by upwelling and entrainment of warmer water from lower parts of the water column is indicated by a point scatter towards lower salinities. Waters near the seabed however, remain largely unchanged (cf. upper right end of final point-cluster in Fig. 2a) because convection did not penetrate through the entire water column. This is confirmed by a temperature synopsis (Fig. 3).

Convection induced by the first atmospheric event caused a far less stable stratification and a general increase in salinity for most parts of the water column. However, no increase in salinity of bottom waters was obtained. The reason for this result is that large parts of the water column had temperatures well above the freezing point. Penetrative convection causes an upwelling of this warmer water (cf. Fig. 3). Once it reaches the sea surface it hampers both ice-growth and brine-release and, hence, energetic haline convection. The negative thermal feedback explains why the existence of a polynya as such need not necessarily imply formation of dense bottom waters. This experiment, in which convection did not penetrate through the entire water column (cf. Fig. 3), is an example for case B in our cartoon of processes (Fig. 1).

In the second experiment a pre-conditioned colder ocean was considered and a distinctly different evolution of water mass formation (Fig. 2b) was obtained. At the end of this experiment much higher salinities were predicted for the entire water column (cf. TS-clusters at 12 and 156h in Fig. 2b). This was caused by a pronounced release of brine from freezing sea ice during the last 70h of the experiment. Since initial temperatures within the water column were still well above the freezing point the remaining heat reservoir had at first to be removed by thermal convection before ice formation commenced. The resulting negative thermal feedback kept the water column virtually ice free for the first 80h of the experiment.

With the onset of thermal convection an internal convective instability was induced by the thermobaric effect. It caused a very rapid internal thermal homogenisation of the water column. This is seen by the transition from a bent TS-cluster at 24h, covering a temperature range of 0.7$^\circ$C, to a much narrower cluster with a range of ca. 0.2$^\circ$C at 48h (Fig. 2b). Further progress in convective thermal homogenisation is characterised by a narrowing of clusters around salinities of 34.783 whilst temperatures approach the freezing point (cf. clusters for 48 and 108h in Fig. 2b).

A point scatter along the freezing line in the TS-cluster predicted after 108h indicates existence of both brines and super-cooled, fresher melting water in the model domain. After 110 simulation hours haline convection prevailed. The entire water column is at the freezing point and, due to substantial brine-release, salinities have increased considerably (cf. results after 156h in Fig. 2b).

This second experiment is an example for case (a) in the cartoon (Fig. 1). It demonstrates that a short-lived atmospheric event may indeed lead to a substantial modification of water mass properties within a polynya once preceding convective pre-conditioning has destabilised the water column and resulted in temperatures close to the freezing point. In the course of the experiment convection shifted from initially thermal to haline buoyancy forcing.
Ice–ocean interactions

An example of ice–ocean interactions from the first experiment (warmer ocean) is given by predicted contours of potential temperature and streamfunction and by the ice thickness (Fig. 3). Results were obtained after 150 simulation hours. Convection caused agglomerations of ice due to convergent surface currents. Agglomerations
of ice generally indicate source regions for negative buoyancy of penetrative convective plumes. Their position changed with a typical time scale in the order of a few hours (i.e. the life time of a convective plume). Downward velocities in funnels of convective plumes (cf. contours of both temperature and streamfunction in Fig. 3) locally exceeded values of 10 cm s\(^{-1}\). Positive anomalies of salinity at the sea surface (not shown) coincide with regions of thin ice where the oceanic heat loss is at maximum (cf. top panel in Fig. 3). Regions of fastest ice-growth account for highest rates of brine-release. They occurred in areas of divergent surface flow where thin ice was permanently formed anew because the neighbouring ice was drawn towards the centre of a surface convergence, i.e. into a convective funnel. This “ice-suction” is caused by a superposition of Ekman ice drift and convergent currents during active convection. The ice-suction is asymmetric and most pronounced upstream of a convergence in relation to the wind drift, i.e. where both motions have the same sign (in Fig. 3 to the left of a funnel). The interactions between convection and ice dynamics appeared to be highly non-linear. Despite high heat losses of the ocean net ice production for this experiment yielded an average ice thickness of only 20 cm after 160 h. This low production rate was caused by the negative thermal feedback referred to earlier. The non-linear interaction between convection and sea-ice causes characteristic transient patterns in the ice cover. Such patterns have been observed in high-resolution SAR (synthetic aperture radar) satellite images of the Greenland Sea in regions where active convection was assumed to occur (Carsey and Garwood,
To investigate ice–ocean interactions depending on wind-forcing the three-dimensional CUBE version of the IOCM (Kämpf, 1996) was applied on a domain with a dimension of 2 by 2 km. Beginning from a state of rest the initial temperature of the homogeneous water column was set close to the freezing point and the same depth as for the SLICE experiments was used. In the first case, in which a low wind speed ($2 \text{ m s}^{-1}$) was applied, ice agglomerations formed cellular patterns (Fig. 4a) similar to classical Bénard cells in free convection. In the second case, where a stronger wind-forcing ($10 \text{ m s}^{-1}$) was applied, frazil and young thin ice agglomerated in undulating streaks closely aligned with the wind direction (Fig. 4b) as found with Langmuir circulation cells. Streaks of young ice aligned with the wind have indeed been observed in recently opened wind-induced latent heat polynyas (E. Carmack, pers. comm., 1995). The separation scale of convective streaks depends on the penetration depth of convection; typical convective aspect ratios in our model experiments ranged between 1 and 3, in accordance with laboratory experiments. The spatial scales of Langmuir circulation, which is not driven by buoyancy forcing, depend on the wavelength of incident surface waves. Possible interactions between Langmuir and convective cells still need further investigation.

The IOCM predictions suggest that distinct and small scale features in ice-patterns are indicative of the state of the underlying oceanic convection. This will depend on both wind-forcing and stratification. Detection of active convective regions by pattern-recognition of remotely sensed scenes of new sea-ice might be feasible. Recent progress in remote sensing, detection of thin ice by passive microwave for instance (Wensnahan et al., 1993), or by high-resolution SAR images (Carsey and Roach, 1994), provides support for this idea.

![Figure 4](image-url)

Figure 4. (a) Ice patterns predicted by three-dimensional IOCM-CUBE after 24 simulation hours. Convection is forced by low winds ($2 \text{ m s}^{-1}$), wind direction aligned with positive y-axis. Mean heat loss: $170 \text{ W m}^{-2}$. Bright areas denote ice thickness in excess of 0.4 m. Model domain 2 by 2 km; isotropic grid size: 10 m.
Gravitational collapse of polynya waters after a convection event

The IOCM model did not produce isolated bodies of cold and salty water masses at the shelf bottom. Only a general increase in salinity for the entire water column was obtained. How can the IOCM results be interpreted in view of observations of isolated lenses of dense bottom water masses (Midttun, 1985) which have been produced by convection?

This apparent discrepancy is explained by processes that take place when the polynya has closed after the passage of an atmospheric event. This was not considered in the convection experiments. The IOCM results suggest to us that a water body with the same spatial scale as the polynya itself undergoes convective water mass formation for the entire water column. This is a result of the very localised atmospheric forcing in the open waters of a polynya. The surrounding water body of the polynya, however, remains comparatively unchanged because the closed ice cover prevents deep reaching convection and water mass formation. Therefore, once the forcing ceases, the entire water column of a polynya, being denser than its ambient waters, will undergo a gravitational collapse, and the surrounding, less dense waters will close over it. Since spatial scales of shelf polynyas are generally larger than the local Rossby radius of deformation we anticipate rotation to influence if not control the gravitation collapse as proposed, for instance, for the Central Bank vortex in the Barents Sea (Quadfasel et al., 1992).

The IOCM results only concerned the initial formation phase of water masses where rotational effects may be neglected because, during active convection, plumes with scales well below the Rossby radius dominate dynamics. The simulation of a gravitational collapse and
the further rotational and frictional motion of an isolated plume of dense water on a structured shelf topography and the final export of water masses beyond the shelf edge requires a different modelling approach.

A numerical experiment was conducted to obtain an estimate for a typical time scale of the re-stratification caused by a rotational gravitational collapse of polynya waters. The initial density contrast to ambient shelf waters was taken from the second IOCM experiment described above. The simulation of the collapse was carried out with a hydrostatic reduced-gravity plume model which predicts the space–time evolution of an isolated body of dense water on an inclined plane shelf topography (a slope of 0.02 degrees was assumed). The plume model (Rubino, 1994) is a two-layer version of the model described by Jungclaus and Backhaus (1994). It was initialised by a cylindrical column of dense polynya water with a diameter of 45 km which covered a water column of 100 m from surface to bottom. For this simple, schematic, and qualitative experiment (applied grid size: 1.5 km) both entrainment of ambient water by the gravity plume and dynamic effects of a large scale shelf circulation were neglected. Snapshots of the simulated collapse of polynya water (Fig. 5) show the slow motion of the dense plume. The centre of the plume only moved about one polynya diameter in 45 days (i.e. 1 km d⁻¹). This is in good agreement with the observed persistence of isolated lenses of dense bottom water on polar shelves (Midttun, 1985; Quadfasel et al., 1992).

Arctic breakouts of cold air occur approximately every 10–20 days. The model experiment on re-stratification due to a gravitational collapse suggests a similar time scale. Hence, pre-conditioning of water masses might indeed take place provided advection by a large scale flow plays a minor role. In that case the frequent occurrence of a polynya in the same region would imply an effective, but localised production of dense bottom water.

Slope convection

One of the few cases of observed slope convection in the Arctic is the outflow of dense bottom waters from Storfjord, Svalbard into the Norwegian Sea (Quadfasel et al., 1988; Schauer, 1995). These observations served as validation for a simulation of slope convection in which a hydrostatic two-dimensional reduced-gravity plume model (Jungclaus and Backhaus, 1994; Jungclaus et al., 1995) was applied. The model predicts the dynamics of a transient bottom-arrested gravity plume of arbitrary shape on a realistic topography. In agreement with observations (Quadfasel et al., 1988) the Storfjord simulation showed that the plume, on its descent of both shelf and continental slopes, entrains different ambient water masses, such as East Spitsbergen shelf water (ESW), Atlantic water (AW), and Norwegian Sea deep water (NSDW). The entrainment of three different water masses led to distinct features in the simulated TS-diagram which compared well with a TS-diagram obtained by Quadfasel et al. (1988). The latter have estimated an increase of the plume size due to entrainment in the order of 500% whereas the model predicted a slightly smaller value of 440%.

A remarkable result of the simulation is that the plume splits at a topographic saddle point at depths around 2.5 m. However, this has as yet not been confirmed by direct observations. One branch of the simulated plume followed a deep trench east of the Knipovitch Ridge which is situated to the west of the Svalbard continental slope. Barents Sea water masses have indeed been found in that area (Gascard, pers. comm., 1994) which we interpret as indirect support for our model result. The other branch of the plume, in agreement with the observations by Quadfasel et al. (1988), proceeded towards the Fram Strait at depths greater than 2.500 m. Slope convection originating from within Storfjord, according to the model, provides dense shelf waters to water mass formation in both the Arctic Ocean (via the Fram Strait) and the Norwegian Sea. An investigation of the thermobaric effect (Jungclaus et al., 1995), which is expected to increase the negative buoyancy and hence, dynamics of an intruding cold plume, yielded a retarded intrusion of the plume at depth. This was caused by increased entrainment of ambient water masses due to enhanced dynamics. Entrainment, however, tends to reduce the density contrast of a plume. The resulting negative feedback of an enhanced entrainment on plume dynamics explained the, at first, surprising result.

On its descent the plume entrains warm AW and it becomes warmer than the NSDW which it encounters at greater depth. This illustrates the already mentioned vehicle effect of initially cold and saline shelf plumes in the Arctic: whenever they entrain warmer and saltier (Atlantic or intermediate) water they account for a downward transport of both heat and salt. This net result of slope convection is in contrast to convection in the open water column, for instance in a shelf polynya or in the Greenland Sea, where convection generally causes an upward transport of both heat and salt (Rudels and Quadfasel, 1991). This also emerged in the polynya experiments described earlier.

Sediments in slope convection

The specific role of sediments in slope convection and water mass formation was considered in a recent investigation (Fohrman, 1996). For that task the plume model was modified to include prognostic equations for two different sediment classes (20 μm and silt with 63 μm). The new model predicts differential (auto-
Figure 5. Simulated gravitational collapse of an initially cylindrical plume of polynya water (height: 100 m, diameter: 45 km) on an inclined topography (slope angle: 0.02°) in a two-layer ocean. Predicted contours of plume height (CI = 10 m) for simulation days as indicated.
suspension and deposition of sediments encountered by a plume on its descent. The additional negative buoyancy due to suspended sediments causes entirely different plume dynamics to those of plumes which are driven solely by an excess buoyancy due to water mass properties, i.e. water mass plumes. Typical velocities predicted within a plume were in the range of 0.1–0.2 m s\(^{-1}\) and 0.3–0.5 m s\(^{-1}\) for a water mass and a sediment plume, respectively. Dynamics may be enhanced in such a way that they are not over-ridden by increased entrainment. This constitutes a remarkable difference to our results on the thermobaric effect. Therefore, in contrast to classical water mass plumes, sediment-laden plumes may descend in an ageostrophic dynamic balance, i.e. their path may intersect depth contours by a large angle.

The enhanced plume dynamics lead to different mixing and entrainment properties and hence, to different water mass formation (Fohrmann, 1996). This is illustrated by predicted TS-diagrams (Fig. 6) for a hypothetical slope convection influenced by suspended sediments. The same ambient water masses were used as for the Storfjord simulation of Jungclaus et al. (1995). It was
assumed that the continental slope was covered with a sediment layer of 1 mm thickness. This layer will be entrained by the plume on its descent. Three cases for a plume with an initial sediment load of 1 g l\(^{-1}\) were considered, the only difference being the initial characteristics of its source water masses which are given in Figure 6. The case of shelf bottom water (SBW) emerging from Storfjord (Fig. 6a), also considered by Jungclaus et al. (1995), is compared with two other hypothetical cases (Fig. 6b,c) where warmer and less saline source water masses were prescribed in order to highlight the role of sediments. For the case in Figure 6(c) the density of the liquid phase of the sediment plume was 27.9 sigma units, which is less than the density of the ambient water masses (28.1 sigma units). However, due to its initial sediment load the plume (liquid phase plus suspended sediment) still had a density of 29.1 sigma units. In these cases studies the comparatively high initial sediment concentration was chosen to ensure a stable stratification also for the third experiment (Fig. 6c).

In all three experiments differential entrainment of the sediment plume yielded a mixing towards NSDW properties as a final result of slope convection. These results were obtained after only three days of simulation. For a water mass plume only the source characteristics of dense (cold and saline) SBW would allow for a mixing towards NSDW properties and a much longer simulation time of about 100 days was needed to obtain this result (cf. Jungclaus et al., 1995). The predicted TS-diagrams (Fig. 6) suggest that the existence of dense shelf bottom waters need not necessarily be a pre-requisite for slope convection if the excess buoyancy of a plume is induced, or at least enhanced, by sediments.

Due to the ageostrophic nature of their dynamics sediment-laden plumes may penetrate deeper and, apparently, much faster than water mass plumes. A gravity plume, when reaching its equilibrium density level, may lift off the bottom to become a lateral intrusion into the adjacent ocean (case f in Fig. 1). Energy and turbulence within the plume would eventually die out which, in case of a sediment-laden plume, would allow sediments to settle. With a diminishing sediment load, however, a plume may become lighter than its ambient fluid which would initiate upward directed convection in the water column originating from the intrusion level. This interesting facet of slope convection has indeed been inferred from observations in the tropics, in the Sulu Sea (Quadfasel et al., 1992), and was later confirmed by a laboratory experiment (Kerr, 1991). However, for the Arctic Ocean the importance of internal convection in water mass formation as a consequence of sediments in slope convection has still to be confirmed. A model experiment conducted for polar water masses (Fohrmann, 1996) showed that a decelerating sediment-laden plume which remains attached to the seabed may indeed become lighter than its ambient waters. This is demonstrated by comparing the predicted actual density of the plume (liquid phase plus suspended sediment) with the density being only a function of predicted water mass properties (Fig. 7). A substantial part of the plume, in particular at its intruding head, will attain a positive buoyancy once the suspended sediments have been deposited. Due to its hydrostatic nature the plume model cannot, however, be applied to simulate the expected upward directed convection. Note the large angle between depth contours and main axis of the plume (Fig. 7) which resulted from ageostrophic plume dynamics.

**Final remarks**

This paper attempts to provide an overview and a description of processes and interactions involved in convective formation and export of shelf water masses in the Arctic Ocean. To achieve a better and more detailed understanding of these processes a suite of process-oriented numerical models was applied. Gaps in spatial scales amongst the models were bridged via data exchange. Scales from both below and above the internal Rossby radius of deformation were considered. A number of uncertainties still remain. For example, because of the lack of direct observations of slope convection hypothetical cases had to be investigated for the intruding plumes which, however, considered typical ambient water masses. Moreover, the existence of upward directed internal convection can only be hypothesised.

Episodic breakouts of cold polar air provide a substantial contribution to convective water mass formation on polar shelves. They cause lee-polynyas which allow for a localised formation of shelf bottom water masses, in late winter in particular after the shelf stratification has been eroded by cooling and ice formation. Simulations of small scale convection and ice formation within a polynya for repeatedly occurring breakouts of cold air yielded oceanic heat-losses in excess of 1000 W m\(^{-2}\). The water column reacts to the atmospheric forcing with a rapid convective formation of water masses whereby its surface instability is at first driven by a thermal and later, with the water column approaching freezing temperatures and so allowing for onset of ice-growth, by a predominantly haline buoyancy flux. Via interactions with the underlying ocean new ice is arranged in characteristic transient patterns, i.e. leads and agglomerations in the form of streaks, which appear to be detectable by modern remote sensing techniques. Ice formation and, hence, convection may be hampered by a negative thermal feedback due to upwelling of warmer water masses. Therefore, only if the entire water column
has been cooled to freezing temperatures may sporadic and short-lived atmospheric events contribute significantly to local pre-conditioning and formation of dense shelf bottom water.

A full validation of the model predictions is impossible at present because suitable process-oriented observations are missing. Simulated ice-patterns however, which are caused by interactions with active oceanic

Figure 7. Predicted density contrast (kg m$^{-3}$) of a sediment plume on a continental slope. (a) Actual density contrast as a result of TS and suspended sediments; (b) density contrast of water masses within plume, ignoring contribution by sediments. Shaded areas indicate regions where positive buoyancy would occur after settling of sediments. Note: large ageostrophic angle between plume axis and depth contours (dashed lines). (Adopted from Fohrmann, 1996.)
convection, resemble features which have already been observed. Furthermore, the model results provide guidance for a strategy of future field observations in shelf polynyas in which the small scales of relevant processes and detection of ice patterns by remote sensing need to be considered.

The gravitational collapse and the slow rotational motion of convectively formed bottom water favour a pre-conditioning for water mass formation in lee-polynyas because the lenses of dense water are retained nearby for long periods of time (months). The export of shelf bottom water by slope convection and the resulting formation of slope water masses due to entrainment has been simulated successfully and validated for the Storfjord outflow. The existence of slope convection in other places in the Arctic Ocean can be inferred from water mass analysis. Its direct observation, however, is hampered by severe environmental conditions and limited accessibility of key regions. Future observations may be guided by additional applications of the plume model.

It is interesting to note that shelf regions in the Eurasian Arctic are much richer in topographic obstacles in exposed positions (islands, archipelagos, headlands, etc.) than the shelves of the North American continent where the Canadian archipelago is blocked by the convergence of the transpolar ice drift. These conditions favour a higher abundance of wind- and tide-induced polynyas in the Eurasian part of the Arctic and hence, convective water mass formation. Consequently, feedback from the ocean to the atmosphere via the release of heat and moisture from polynyas is to be expected primarily in the Eurasian shelf regions of the Arctic. Meteorologists have complained about the atmosphere being too dry in their models of the Arctic with the result of the baroclinicity being too weakly represented. They have realised that episodic releases of oceanic heat and moisture on small spatial scales need a more realistic parameterisation in their models and our results will be of value in this context.

The potential role of sediments in slope convection and water mass formation in the Arctic Ocean has not been considered in much detail as yet. However, substantial amounts of recent sediments of either glacial or fluvial origin are available primarily as with the occurrence of lee-polynyas, in the Eurasian shelf regions. Slope convection in the Arctic Ocean enhanced by suspended sediments appears, therefore, to be another issue worth further investigation that should include the study of mechanisms that contribute to the re-mobilisation of sediments for slope convection.

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