ICES Journal of Marine Science, 54: 351–365. 1997

Water mass transformation in the Barents Sea – application of the Hamburg Shelf Ocean Model (HamSOM)

Ingo H. Harms

Harms, I. H. 1997. Water mass transformation in the Barents Sea – application of the Hamburg Shelf Ocean Model (HamSOM) – ICES Journal of Marine Science, 54: 351–365.

The bottom water formation in the Barents Sea plays an important role in the maintenance of the Arctic halocline. Deep water is partly formed by haline convection caused by brine release during ice growth. To investigate this process, the HAMburg Shelf Ocean Model (HAMSOM) was coupled to a dynamic and thermodynamic ice model and applied to the Barents and Kara Seas. The coupled model is initialised with climatological temperature and salinity data and forced with realistic wind stresses and air temperatures.

The results of simulated winter scenarios show that the role of the thermo-haline convection is two fold. First, convection in the upper layers erodes the locally pronounced haline stratification while ice is formed in major portions of the area. Second, local and small-scale openings in the ice cover (polynyas) provide the necessary brine release for dense bottom waters. The polynyas open frequently near Franz Josef Land and Novaya Zemlya due to off-shore winds. Tidal currents, however, may also play a role in opening the closed ice sheet. Tracer studies reveal that most of the bottom water leaves the Barents Sea through the Svyataya Anna Trough towards the Arctic Ocean. Outflow rates of the deep water as well as the temporal and spatial evolution of stratification and ice cover are presented.

Key words: Arctic, Barents Sea, water mass transformation, modelling.

Received 6 March 1996; accepted 1 February 1997.

I. H. Harms: Institut für Meereskunde, Universität Hamburg, Troplowitzstr. 7, D-22529 Hamburg, Germany.

Introduction

The formation of ''Shelf Brine Water'' (SBW) is a result of brine release during intense ice formation in shallow Arctic shelf regions. The dense water accumulates at the bottom and flows via deeper troughs towards the shelf margins where it loses its initial density due to entrainment with fresher ambient water masses (Rudels and Quadfasel, 1991). Finally, the bottom waters descend at the continental slope and penetrate into the deeper layers below the Arctic halocline. The production and outflow of SBW has a strong influence on the vertical stratification of the Arctic Ocean because it forms an important contribution to the maintenance of the Arctic halocline (Aagaard *et al.*, 1981). This makes the SBW production on Arctic shelves of vital importance in regard to both climate change and arctic sea-ice variances: a decrease in the production of shelf brine water weakens the isolating effect of the Arctic halocline and brings the sea-ice in contact with warmer Atlantic water masses, affecting the Arctic sea-ice cover.

The presence of highly saline bottom water on Arctic shelves has been recognised for many years (Nansen, 1906) even though direct observations are scarce due to severe climate conditions particularly in winter. Latest observations reveal that intensive bottom water formation takes place in the north-eastern Barents Sea near Novaya Zemlya and Franz Josef Land (Midttun, 1985) but also near Svalbard (Quadfasel *et al.*, 1988; Blindheim, 1989; Schauer, 1995). The model study described in the following section focuses on the large-scale interaction of atmospheric cooling, ocean circulation and ice formation resulting in a transformation of water masses. The formation of SBW is usually linked with the process of convection which, unfortunately, needs to be parametrised in numerical models using hydrostatic equations. The present model simulations, therefore, do not refer to the physical process of convection but to the detection of convective events and to the outflow of bottom water from the shelf on a basin-wide scale. The results are part of a combined model study on convection and Arctic bottom water formation which

1054–3139/97/040351+15 $25.00/0/jm970226

also encompasses small scale and process-oriented modelling (Backhaus *et al.*, in press).

The Barents and Kara Seas general circulation model

The coupled ice-ocean general circulation model (GCM) of the Barents and Kara Seas (Fig. 1) is based on the Hamburg Shelf Ocean Model (HamSOM) which is a three-dimensional, baroclinic, circulation model, developed at the Institute of Marine Research (University of Hamburg) for investigations of shelf sea processes (Backhaus, 1985). The model uses non-linear primitive equations of motion invoking the Boussinesq approximation. Furthermore, the hydrostatic approximation and the equation of continuity are applied to predict the elevation of the free surface from the divergence of the depth mean transport. The numerical scheme of the circulation model is semi-implicit which allows for economic time steps. The equations are discretized as finite differences on an Arakawa C-grid. A detailed description of the circulation model with recent applications to the coastal waters of Canada (Vancouver) can be found in Stronach *et al.* (1993).

The circulation model is coupled to a thermodynamic and dynamic sea-ice model in order to simulate the surface fluxes for temperature and salinity. The applied one-layer ice model was kept simple; it ignores an ice rheology (free drift) and snow cover but predicts space and time dependent variations of ice thickness and compactness (Hibler, 1979). At open boundaries a zero-gradient boundary condition is applied for both variables. Thermodynamic processes are based on well-known numerical methods first introduced by Maykut and Untersteiner (1971), Semtner (1976) and Parkinson and Washington (1979). Ocean and ice models communicate via fluxes of momentum, heat and salt:

- the sea surface temperature depends on surface heat fluxes, calculated with standard bulk formulae (Maykut, 1986);
- sea surface salt flux (brine or freshwater release) is proportional to the volumetric thermodynamic ice growth (Lemke *et al.*, 1990);
- transfer of momentum from the atmosphere to water depends on the fractional ice cover (compactness) and the ice drift; and,
- thermodynamic ice growth is determined from heat flux balance equations for the top and bottom of the ice cover (Parkinson and Washington, 1979).

Convective overturning is parameterised by a combination of vertical swapping and mixing of instable layers. The model allows for the simulation of the water mass transformation as a result of surface fluxes (winds, cooling, ice formation and brine release) and the basin-wide advection and mixing of water mass properties. For more information on the Barents and Kara Sea GCM see Harms (1992, 1994).

Ice formation and polynyas

The GCM was driven by six-hourly atmospheric fields of winds, air pressure, and temperature from the European centre for medium-range weather forecasts (ECMWF). Additionally, the M_2-tidal forcing and monthly mean river runoff rates from four major rivers (Pechora, Ob, Yennisey and Pyasina) were included (Harms, in press). The model was initialised with climatological means of temperature and salt (Levitus, 1982) which unfortunately are very poor in Arctic regions. Prognostic simulations for realistic winter scenarios (September–March), however, revealed that the model was able to overcome this disadvantage. Compared to the initial fields the obtained sea surface temperature and salinity distributions are rather complex and finely resolved. This holds in particular for the Barents Sea where the combination of inflowing warm Atlantic Water and atmospheric cooling leads to a distinct Polar Front near Bear Island and in the central parts of the sea as shown by the simulated sea surface temperatures in Figure 2.

A time series of ice cover and ice thickness from September 1988 to March 1989 (Fig. 3) reveals that the ice formation starts as early as September. During the first winter months there is a constant increase of average ice thickness and ice coverage. At the end of December, however, both variables tend to reach a stationary state which means that in late winter no further significant increase of ice coverage or ice thickness occurs. In spite of apparently stagnating ice formation in late winter, there is still a significant thermodynamic ice production in polynyas around Franz Josef Land and Novaya Zemlya. In our model simulations, polynyas appear, for example, at the end of December 1988 and January 1989 (Fig. 4). This type of lee-polynya or ‘‘flaw lead’’ is caused by offshore winds and could last for several days. Even if in our simulations polynyas do not appear as totally open waters but as thin ice areas, the model polynyas produce a significantly higher heat loss to the atmosphere than surrounding regions where the closed and thick ice cover has a strong isolating effect on atmosphere–ocean boundary fluxes.

In reality, the differential cooling is even more striking because polynyas usually appear as totally open water areas, forming a kind of heat-loss window for the isolated ice-covered ocean. The virtually produced thermodynamic ice thickness in polynyas may be in the range of 10 to 15 m during a winter and the corresponding salt release is in the order of 10^{11} kg,

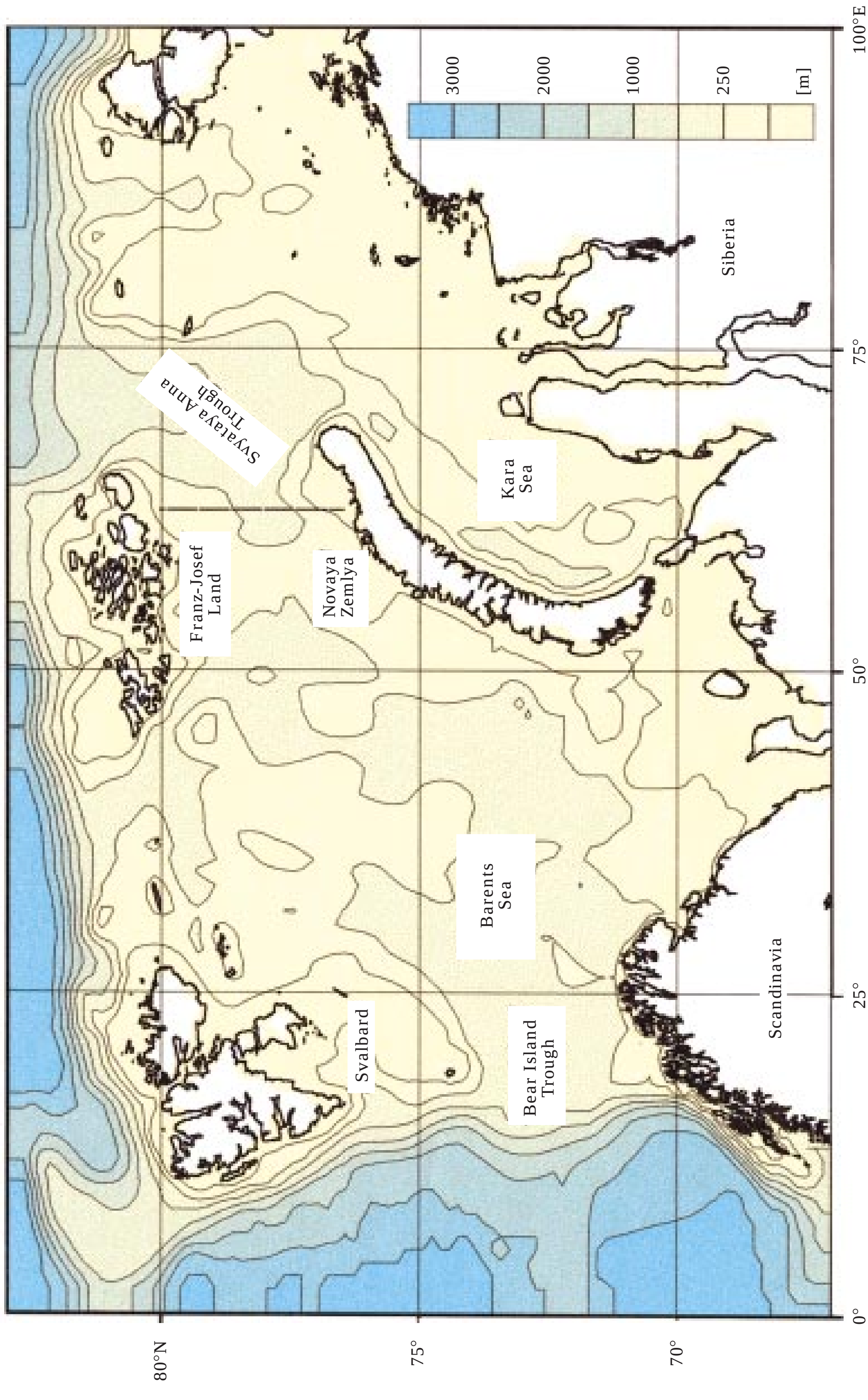

Figure 1. Domain and topography of the Barents and Kara Sea GCM.

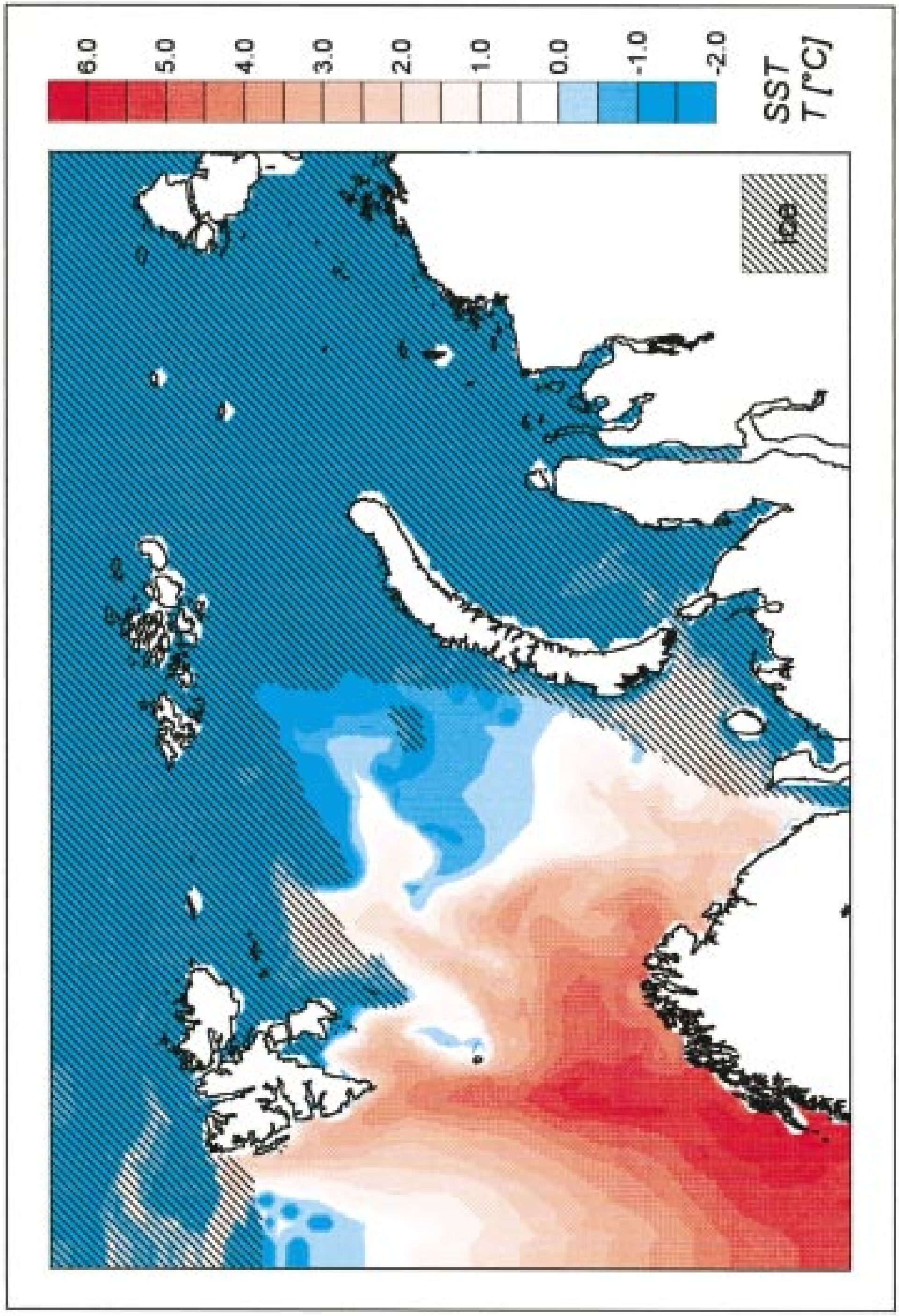

Figure 2. Mean winter sea surface temperatures predicted by the Barents and Kara Sea GCM.

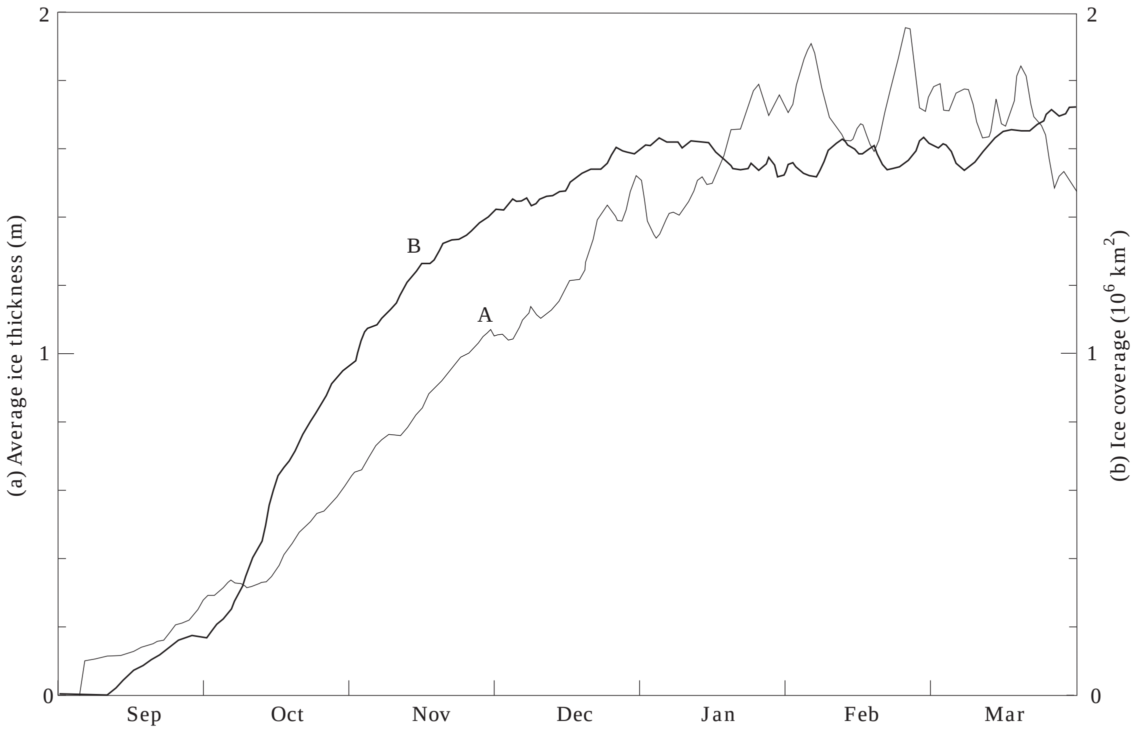

Figure 3. Time series of (a) ice thickness and (b) ice coverage in the Barents and Kara Sea predicted by the GCM for September 1988 to March 1989.

depending on the size of the polynya (Martin and Cavalieri, 1989).

Water mass transformation and ''brine tracer'' releases

The transformation of water masses is depicted by means of vertical profiles for temperature and salinity (Fig. 5) at 79°N and 63°E, close to Franz Josef Land. In October both profiles show a pronounced stratification with less saline water at the top. The fresh surface layer is the result of summer melting and freshwater input. The surface temperature, well above 0°C at the beginning of the simulation (September), has been cooled down to −1.8°C (i.e. the respective freezing temperature). In the following months the temperature and salinity profiles are homogenised by convection resulting in an almost neutral state of stratification at the end of December. From this point on deep haline convection can take place if an intense thermodynamic ice formation provides a sufficient release of brine.

The salinisation of the bottom water is clearly visible in temperature and salinity (TS) characteristics of the deep layers in the Svyataya Anna Trough which is the outflow region of the deep water (Fig. 6). At the different stages during the winter scenario an increase of the bottom salinity only occurs from January to March. The TS-characteristics in the months before are more or less unaffected apart from a slight cooling.

SBW can be traced on a section between Franz Josef Land and Novaya Zemlya. Figure 7 shows a comparison between (a) model results and (b) observations (Midttun, 1985) at the entrance of the Svyataya Anna Trough. Despite the different seasons the agreement between both is remarkably good. The salinity at the bottom is clearly above 34.8 and partly above 35.0. The saline bottom water flows in a north-easterly direction through the Svyataya Anna Trough towards the Arctic Ocean. A classification of the computed Svyataya Anna Trough outflow, with respect to the salinity of bottom waters (Fig. 8), reveals that the outflowing deep water with salinity >34.80 does not pass through the section until the middle of February. Higher saline waters pass through even later, at the beginning or end of March.

In order to locate areas where convection occurs and to trace the SBW, passive tracers were released at the surface in proportion to the brine release. In early winter high amounts of tracers were usually found at the surface layers of the northern Barents Sea due to intense ice formation and a very stable stratification. In late winter, however, the opposite situation occurs. Figure 9 shows the concentration of passive ''brine tracers'' at the surface by the end of March 1989: the total amount of

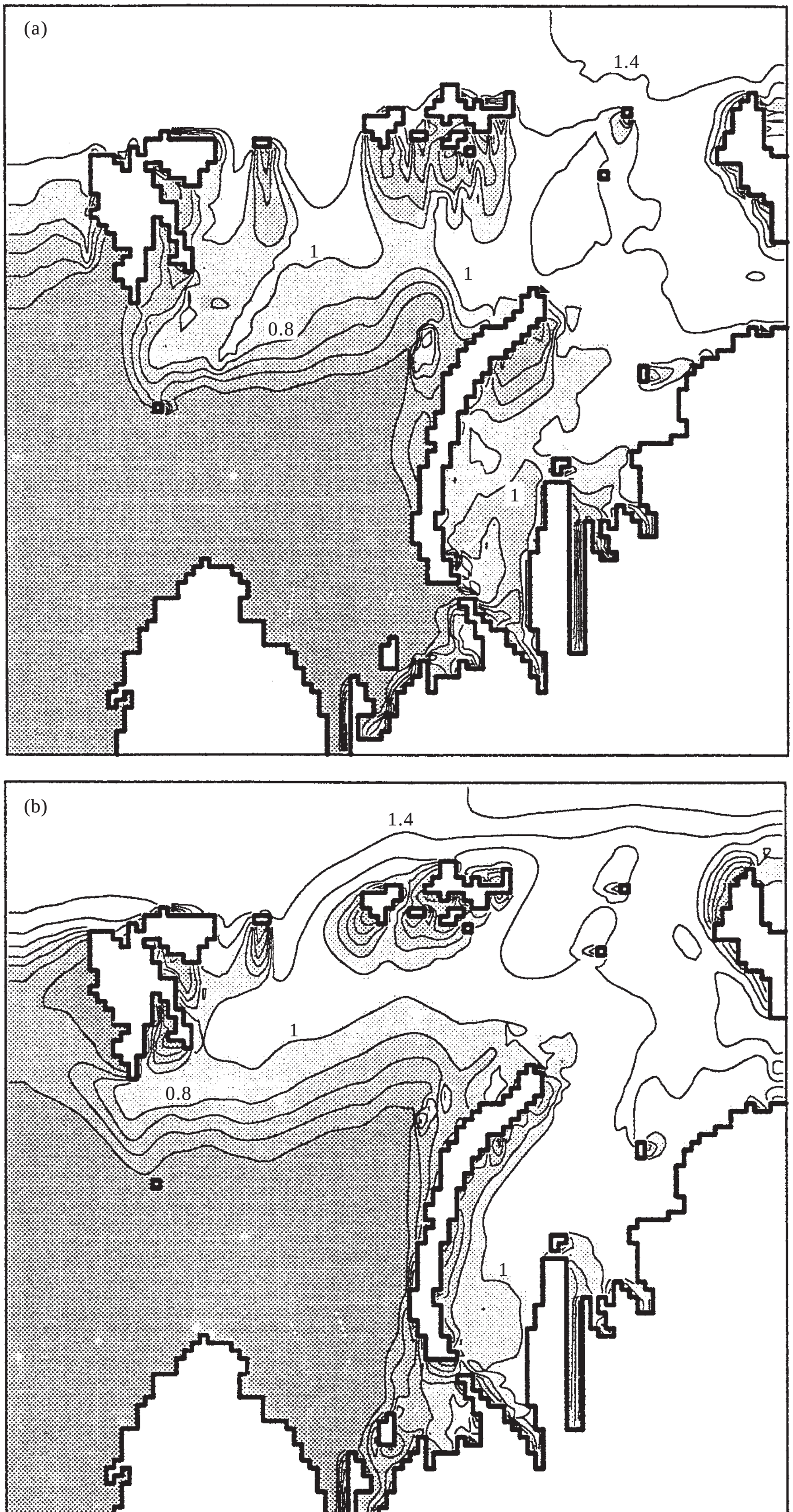

Figure 4. Contours of the mean ice thickness (CI = 0.2 m) predicted by the GCM for the end of the months (a) December 1988 and (b) March 1989.

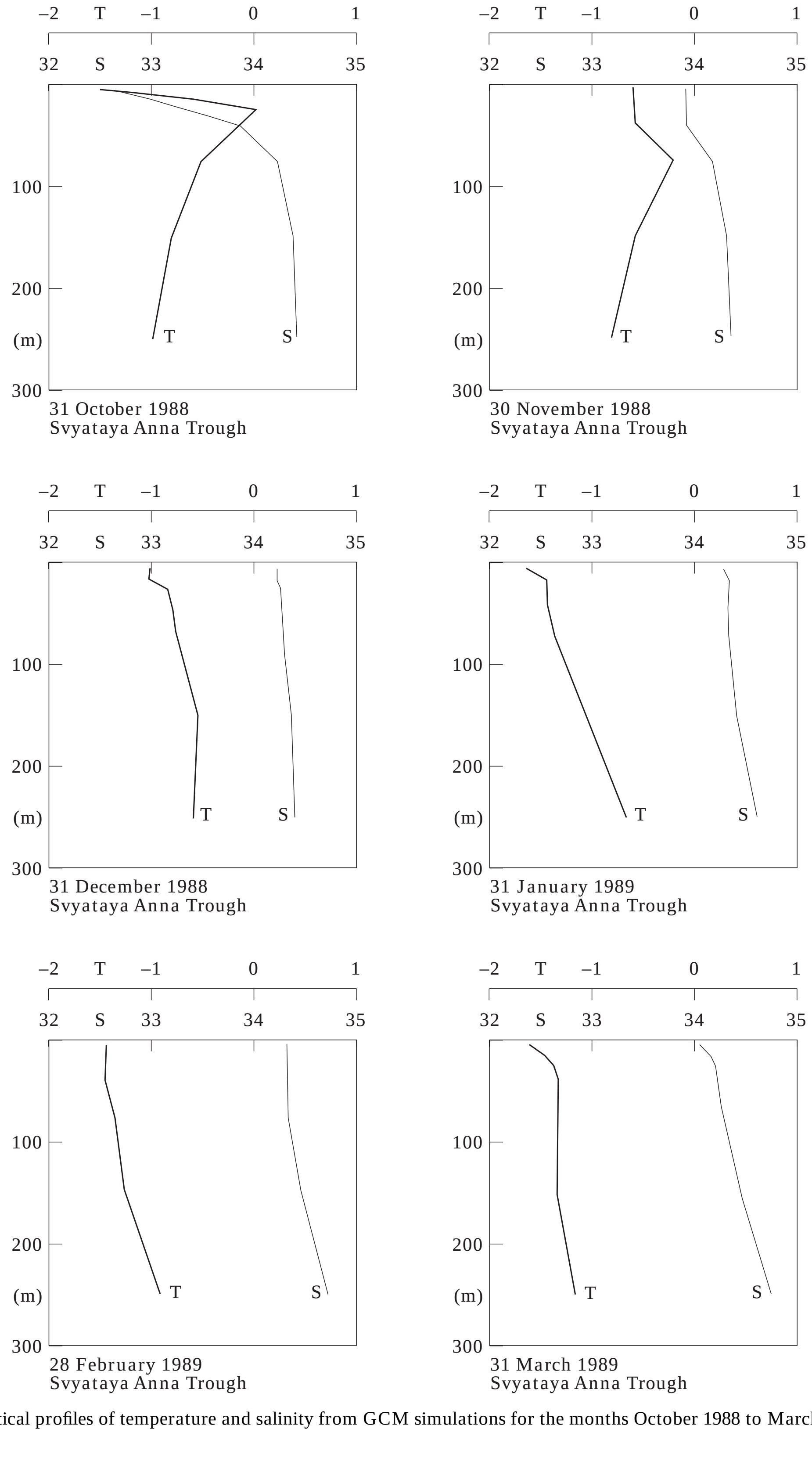

Figure 5. Vertical profiles of temperature and salinity from GCM simulations for the months October 1988 to March 1989 at 79°N and 63°E.

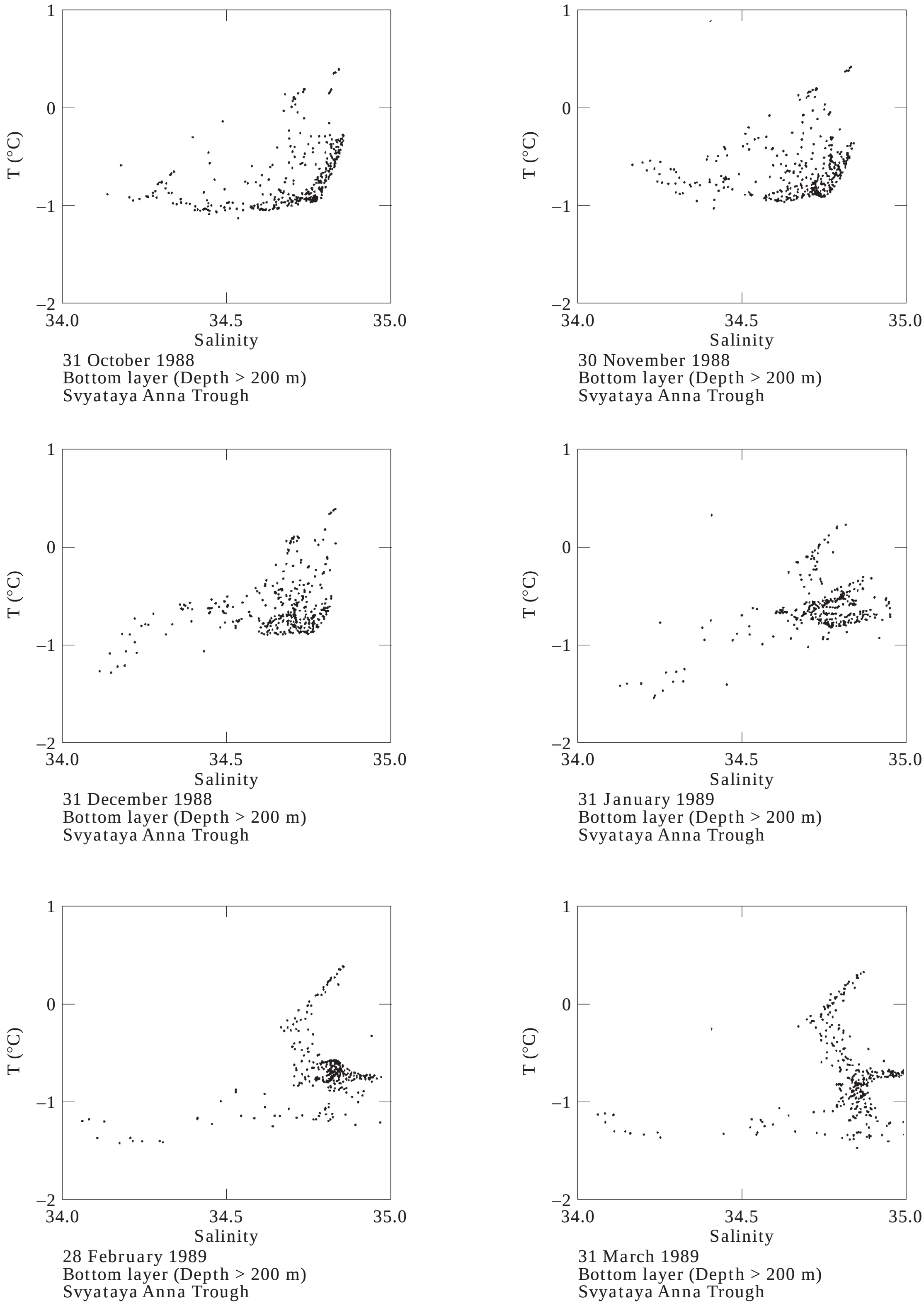

Figure 6. Temperature/salinity diagrams from GCM simulations for the months October 1988 to March 1989 showing the bottom water masses of the Svayataya Anna Trough.

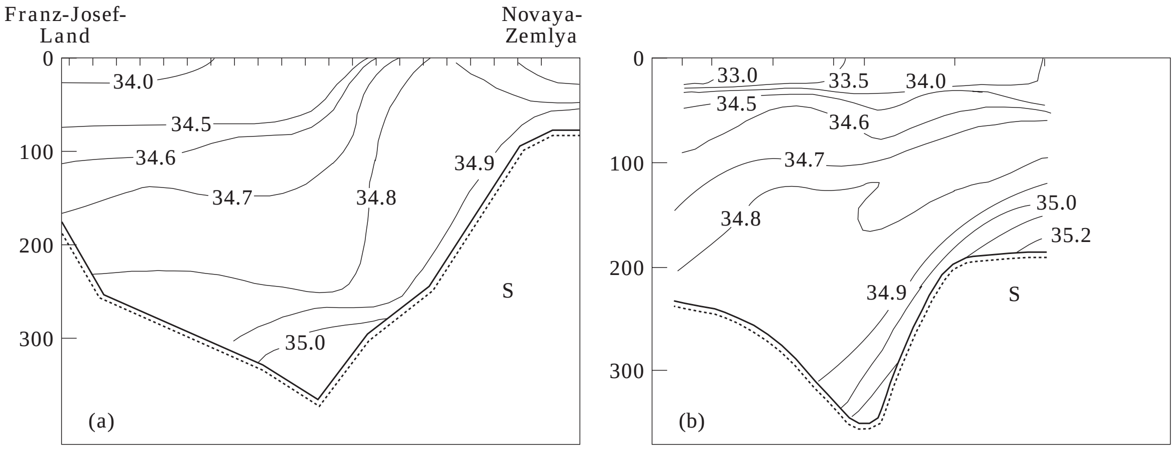

Figure 7. Vertical section for salinity between Franz Josef Land and Novaya Semlya (cf. dashed line in Fig. 2). Comparison between GCM results (a) March, and observations (b) September 1979 (Midttun, 1985).

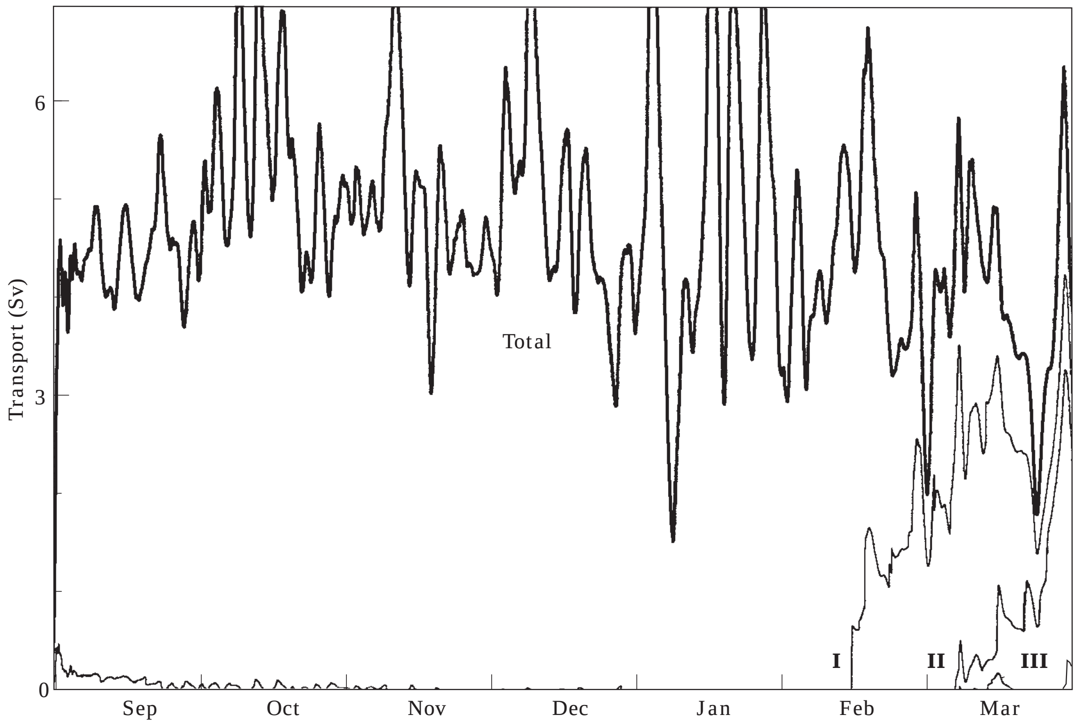

Figure 8. Time series of the predicted shelf water mass export through the Svyataya Anna Trough towards the Arctic Ocean in Sv (1 Mio m^3/s). Thick line denotes the total outflow; thin lines denote transport of water masses that with salinities greater than excess 34.8 (I), 34.85 (II) and 34.9 (III).

tracers in this figure is of minor importance since it gives only a qualitative view of the maximum abundance. The white areas are almost free of tracers whereas the (dark) shaded areas denote enhanced tracer input or brine release. Low tracer concentrations in Figure 9 reveal that in late winter the tracer input into the surface is small compared to the beginning of the freezing period because of a stagnating thermodynamic ice formation. A second reason for the absence of tracers is ongoing convection, which continuously mixes surface tracers from previous months into deeper layers. The tracer input into the surface waters around Svalbard, Franz Josef Land and Novaya Zemlya, exactly where lee-polynyas occur, remains very high

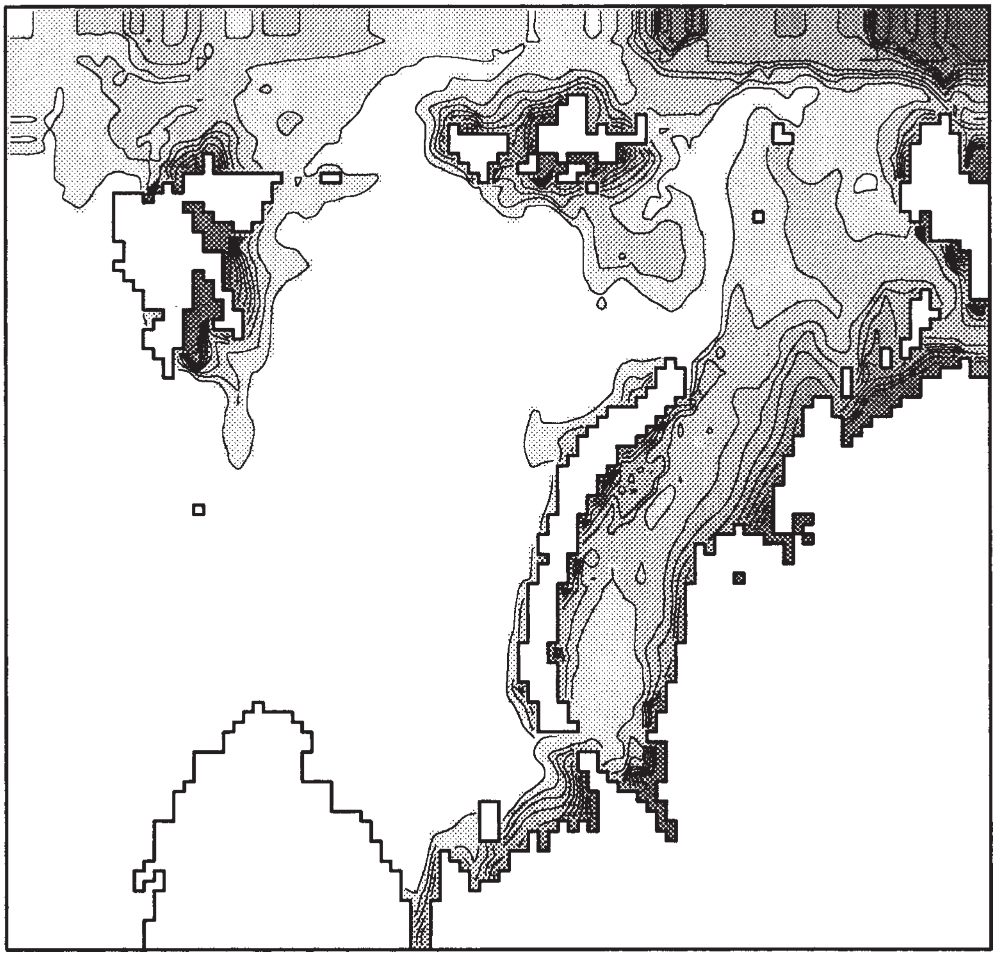

Figure 9. Concentrations of passive ‘‘brine tracers’’ at the surface, predicted by the GCM for the end of March 1989. White areas denote absence or low concentration of tracers, shaded areas denote high tracer concentrations (i.e. strong brine release).

which indicates enhanced brine rejection and ice formation.

As well as Franz Josef Land and Svalbard, the Storfjord and Hinlopen Strait entrances show a high ‘‘brine tracer’’ input. The SBW outflow from these regions goes either westward between Bear Island and Svalbard or northward between Svalbard and Franz Josef Land. The east side of Novaya Zemlya and the Siberian coast also reveals high tracer concentrations. The latter one is, in this context, of minor importance because low surface salinity does not provide enough brine release for bottom water formation.

In general one can assume that the tracer input indicates enhanced local ice formation, brine release and convection in the coastal waters of Svalbard and Franz Josef Land. The model simulations suggest that large amounts of SBW are formed in polynyas near these archipelagos.

The sensitivity of simulated bottom water formation

A considerable uncertainty in our GCM simulations arises from the initial climatological temperature and salinity field. The ice formation and bottom water production results from the model, depend not only on the applied meteorological forcing but also on the initial conditions given by temperature and salinity. This means that the results emerging from different winter scenarios remain very similar as long as they use the same initial fields.

The climate of the Barents Sea (i.e. the combination of temperature, salinity and ice) depends on imported North Atlantic properties. Ådlandsvik and Loeng (1991) reported significant temperature variations on standard sections in the Barents Sea which correlated well with the maximum winter ice extension (i.e. the ‘‘ice index’’) and the Atlantic inflow between Bear Island and the north coast of Norway (Flugløya). A remarkable event in this context was the severe winter 1979/1980 when the ice extension in the Barents Sea reached a maximum value (Loeng, 1991). The ice index from this period is correlated to low temperature and salinity anomalies imported from the North Atlantic/Norwegian Sea.

The spectrum of observed cold and warm climate periods is wide, but a range of 2–4 years is dominant. The dominance of these time spans corresponds with the flushing times for the Barents Sea. The computed GCM flushing times based on monthly mean circulation patterns vary between 2.5 and 3.8 years confirming the high volume flux through the Barents Sea (Loeng *et al.*, 1995). As a highly advective region with a considerable potential for transforming water masses, the Barents Sea is incapable of simply masking climate variation for a couple of years, as, for example, is apparent in the Baltic Sea. It is likely that the observed climate fluctuations in the Barents Sea reflect large-scale variations in the ocean–atmosphere system of the Arctic and North Atlantic.

An important point in this context is the question: ''How sensitive is the transformation of water masses by cooling and ice formation to observed climate fluctuations?'' In order to investigate the sensitivity of our model simulations on this point we varied the initial conditions to simulate climate fluctuation. The winter 1988/1989 represents a mean situation in the Barents Sea. It was a rather neutral winter in a transition time between a cold and a warm period (Ådlandsvik and Loeng, 1991). Therefore, the simulated winter scenario 1988/1989 serves as an ideal control case to investigate the sensitivity of our results in respect to varying initial conditions.

In four sensitivity studies the initial temperature and salinity field was modified and the results were compared to the control run (winter 1988/1989). In the first two studies the initial surface temperatures of the upper 50 m were one degree higher and one degree lower respectively than in the control run. This roughly corresponds to observed temperature variations on standard sections (Loeng, 1991). Ice formation and bottom water production in these two studies did not differ significantly from the control run. The heat loss in the Barents Sea is remarkably effective in that the cooling of the surface waters down to the freezing point and the onset of ice formation appear to be almost invariable from the applied modifications. Very different results emerge from the two other sensitivity studies in which the initial surface salinity was altered. In the first sensitivity run, called (b), the salinity <34.0 of the upper 50 m was increased by roughly 0.5. In a second run, called (c), the upper layer salinity was decreased by 0.5. These modifications only affect the polar surface waters: the initial haline stratification is stronger in (c) and much weaker in (b), without affecting the deeper Atlantic water masses >34.0. The amount of fresh water needed for this modification is in the range of 1200 km^3 which approximately equals the total annual freshwater runoff into the Arctic Ocean.

The different initial haline stratification leads to significant differences in bottom water production and outflow. Compared to the control run the initial weak haline stratification produces much higher outflow rates of saline bottom water than the initial stronger stratification (Fig. 10). In the control run (a) about 50% of the total outflow at the end of March is formed of water masses with salinity >34.85. In case (b) however, these rates increase up to 80–90% whereas in case (c) the rates decrease to less than 10%. These results suggest that the initial haline stratification in late summer significantly affects the bottom water production in the following winter.

Compared to reality the suggested modification in salinity is considerable and probably unrealistic in terms of freshwater input from rivers. However, the haline stratification in Arctic Shelf Seas is not only a result of strong river runoff but also of ice melting in summer which overlays the imported saline Atlantic waters. In the Barents Sea, in particular, the stratification depends more on the Atlantic influence with ice freezing or melting, rather than on the freshwater runoff. It is the combination of both ice melting and Atlantic inflow that creates the pre-condition for the transformation of water masses in winter. Based on our results it would appear that a warm period, high melting rates in summer, strong Atlantic inflow, tends to amplify the stratification in summer which reduces the bottom water production in winter and vice versa.

''Tidal polynyas''

The present study confirms the importance of polynyas and leads for the formation of deep water. Smith *et al.* (1990) gave an overview of the various mechanisms which may create these features. In addition to the ''classical'' mechanisms such as wind, we would like to focus on another possible mechanism for opening a closed ice cover, the tides. Previously F. Nansen (1906) has written about a periodical opening and ridging of the pack ice during his overwintering with the ''Fram''. This was obviously due to diurnal tides such as the K_1, an important tidal constituent in the Arctic Ocean (Kowalik and Proshutinsky, 1993).

Tides may be responsible for enhanced ice movement not only in the open ocean but also near the coast. Near fjord-type shore lines, tidal resonance may lead to considerable elevations. This is particularly true for Svalbard and the Storfjord where amplitudes of up to 70 cm occur (Harms, 1992; Gjevik *et al.*, 1994). Strong

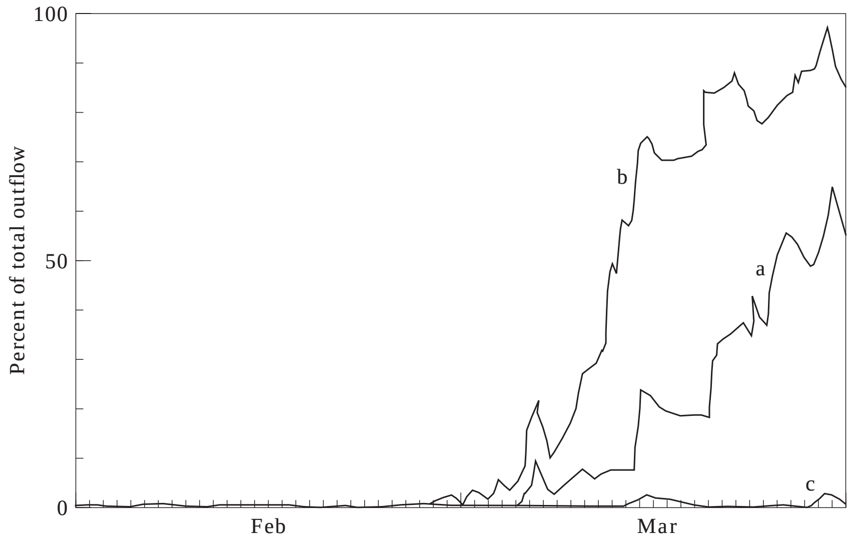

Figure 10. Export of saline bottom water masses that exceed 34.85, between Franz Josef Land and Novaya Zemlya from February 1989 to March 1989. The lines denote the percentage of the total outflow: control run (a), weak initial haline stratification (b) and strong initial haline stratification (c).

tidal elevations are mostly related to enhanced tidal currents which may be responsible for eroding a closed ice cover. Grounded ice over a rugged and shallow topography or landfast ice near a cleft shore line can cause the destruction of a newly formed ice sheet.

An interesting feature in respect of this is the presence of two small throughflows which connect the inner Storfjord with the eastern Svalbard waters. A phase lag of the M_2-tide between the inner and the outer parts leads, for certain time spans, to significant differences in the water levels (Fig. 11). The resulting strong tidal currents within these narrow channels may cause a periodical break up of the closed ice cover. A similar phase lag occurs through the Hinlopen-strait which separates the island ''Nordaustlandet'' in the north-east from the main land of Svalbard. The results of tidal phase lags on a small spatial scale are described dramatically in the Arctic Pilot (Norwegian Polar Research Institute, 1990). Concerning the Sounds that link the Storfjord to eastern Svalbard waters it is written: ''There are strong tidal streams in these waters, especially in the Sounds. . . . Mariners are particularly warned against the Sounds between Spitsbergen and the north side of Barentsoya, where very strong tidal streams combined with pack ice can be a great hazard to the safety of a vessel.'' A more detailed description reads: ''In the narrowest part (of the throughflow), the tidal stream has been measured by 8–9 knots. The Sounds must not be passed when there is much pack ice in them. Even with large engine power, there is the risk of becoming fast in the ice which is pressed together in the funnel shaped areas at the entrances to the Sounds. The tidal streams change quickly in the Sounds, as there are no slack periods worthy of mention at the change of stream.'' Concerning the Hinlopen-strait it is said: ''The tidal stream in Hinlopen-strait is strong, and it runs along the line of the Sound. . . . Pack ice can be troublesome, particularly along the E-side of the Sound. According to whalers and other explorers, there is most usually open water along the W side even though there is otherwise close ice.''

These processes might only have local effects on the ice cover but they happen regularly. It seems plausible that the ice, not only within these channels but also at their entrances, is continuously moving and frequent openings in the pack ice, like small ''tidal'' polynyas, are created. Within these openings new ice is quickly formed during freezing conditions. The associated brine release may contribute, for example, to the well-known bottom water production of the Storfjord (Quadfasel *et al.*, 1988) and might explain why the comparatively small fjord area is able to produce large amounts of saline bottom water. Recent observations show that the outflow of shelf brine water out of the Storfjord is very persistent over more than 130 days in the range of 0.16 Sv (Schauer, 1995).

The situation in Svalbard is repeated in the archipelago of Franz Josef Land, where sea level differences of the order of 20 cm occur. However, descriptions of tidal currents through these islands could not be found. It is suggested that the trapping of tidal waves, a very frequently observed phenomena in the Arctic, leads to enhanced tidal elevations near islands or sea-level elevations. The tidal phenomena described above are therefore not only confined to the Spitsbergen area but might also occur in other appropriate places within the Arctic.

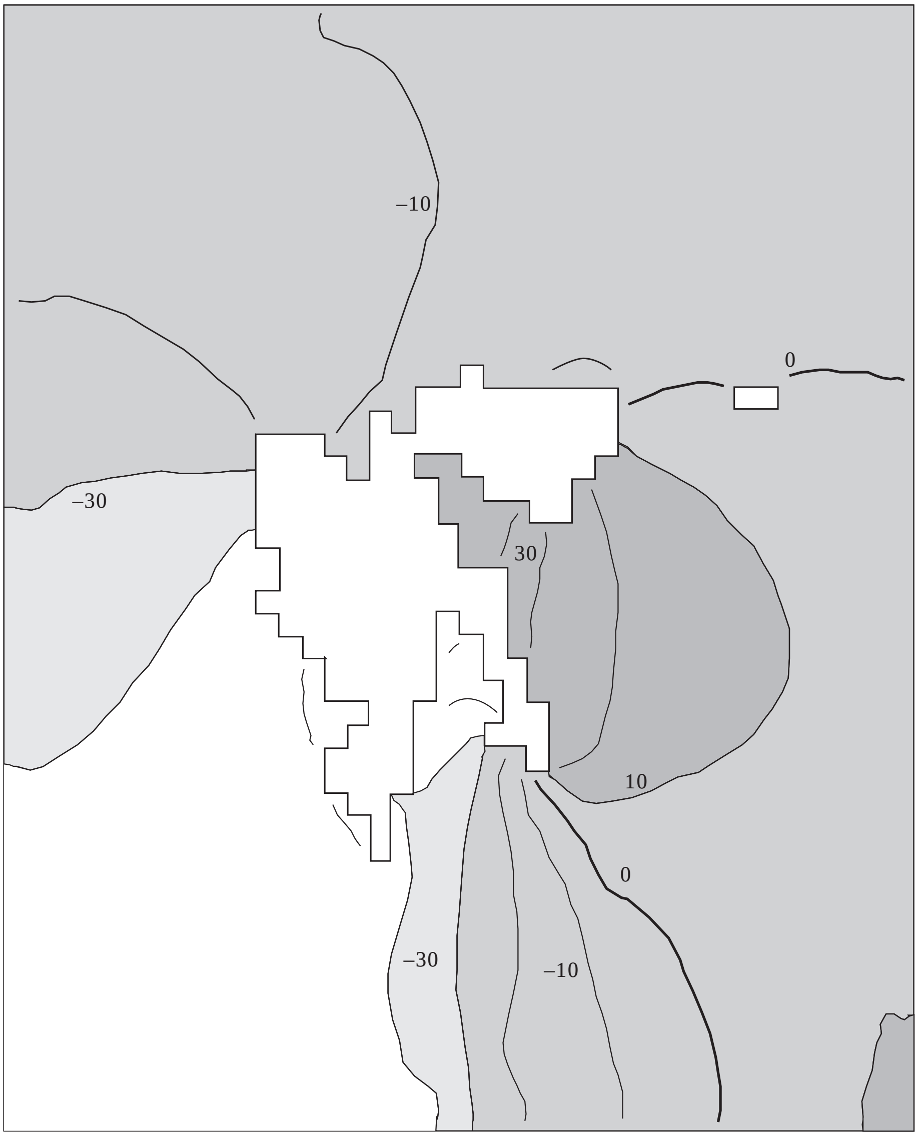

Figure 11. Sea surface elevation (cm) of the M_2-tide around Svalbard predicted by the GCM, 5 hours past Greenwich (CI = 0.1 m).

In a numerical model study concerning the Arctic Ocean tides, Kowalik and Proshutinsky (1994) stressed the interaction between tidal motions and the ice cover. The transfer of tidal water motion to the ice and the non-linear ice flow interaction can be the source of enhanced residual ice motion near topographic irregularities or the continental slope. In certain areas in the Laptev Sea the simulated residual ice motion close to the shore or shore-fast ice is directed away from the coast towards the sea causing open leads of polynyas. The estimated rate of ice production due to the tides is significantly higher along the continental slopes and near certain islands. The authors argue that besides wind effects and the upwelling of warmer (Atlantic) waters from below the tides might be responsible for the opening of the ''Great Siberian Polynya'', a frequently observed open lead in the Laptev Sea (Dethleff, 1995). At present we can only hypothesise that tidal induced openings in the otherwise closed ice cover contribute to the formation of SBW. Further observations of these very local effects are needed to confirm this assumption.

Summary and conclusions

The Barents and Kara Sea GCM reproduces reasonably well, despite starting with relatively poor climatological temperature and salinity data, the dominant hydrographic features produced by atmospheric forcing, ice formation and the advective heat transport from the Norwegian Atlantic Current. The hydrography of the north-eastern Barents Sea at the beginning of the freezing period is characterised by a strong haline stratification which results mainly from ice melting during summer. This stratification inhibits deep haline convection. The onset of ice formation and brine release in autumn weakens the strong haline stratification through convective mixing of the upper layers. Then, in late winter, a homogenised water column is found which allows haline convection down to the bottom. This ''deep haline convection'' requires intensive brine release and therefore high thermodynamic ice production rates. The applied ''brine tracers'' reveal that, due to a general decrease in ice formation in late winter, only polynyas are able to provide sufficiently large thermodynamic ice growth. As well as Svalbard and Novaya Zemlya, Franz Josef Land seems to be a key region for the formation of SBW.

Apart from wind-induced openings in the ice, it can be argued that strong tidal currents might also be an effective mechanism to break up the landfast ice sheet in Storfjord/Svalbard or around Franz Josef Land. This leads to the assumption that ''tidal polynyas'' might also contribute to the formation of SBW.

The outflow of saline bottom water occurs mainly through the passage between Franz Josef Land and Novaya Zemlya into the Svyataya Anna Trough. Flux estimations for the winter 1988/1989 show that water masses with salinities >34.8 leave the Barents Sea in February and March and form up to 70% of the total outflow. Bottom water with higher salinity (>34.9) can be found only in March with transport rates not more than 10% of the total outflow. SBW formed near Svalbard tend to leave the shelf regions to the west between Bear Island and Svalbard or to the north between Svalbard and Franz Josef Land.

Concerning the sensitivity of SBW production and outflow we conclude that the initial haline stratification in late summer is of major importance for formation of SBW in the following winter. If the initial vertical summer stratification is too strong, the deep-reaching haline convection and the formation of SBW in winter might be blocked, while a weak initial stratification causes enhanced bottom water production. The model results were rather independent of initial temperature variances.

The Barents and Kara Sea GCM is able to show that the hydrography and the climate of the region depends greatly on the inflowing Atlantic water masses (warm and saline), the atmospheric heat fluxes and ice formation. Any variation in one of these components will inevitably have consequences for the production of Arctic shelf brine water.

Acknowledgements

This work is part of a PhD thesis, written at the Institut fuer Meereskunde, Universitaet Hamburg. The author is indebted to his advisor Prof Jan Backhaus and to several colleagues for numerous discussions during the course of the model application. Thanks to Jennifer Verduin who carefully checked the manuscript. The work was funded by the Deutsche Forschungsgemeinschaft (DFG-SFB318).

References

Aagaard K., Coachman, L. K., and Carmack, E. 1981. On the halocline of the Arctic Ocean. Deep Sea Research, 28: 529–545.

Ådlandsvik, B., and Loeng, H. 1991. A study of the climatic system in the Barents Sea. Polar Research, 10: 45–49.

Backhaus, J. O. 1985. A three-dimensional model for the simulation of Shelf Sea dynamics. Deutsche Hydrographische Zeitschrift, 38: 165–187.

Backhaus, J. O., Fohrmann, H., Kämpf, J., and Rubino, A. In Press. ''Formation and Export of water masses produced in Arctic Shelf Polyugas–process studies of oceanic convection''. ICES Journal of Marine Science.

Blindheim, J. 1989. Cascading of Barents Sea bottom water into the Norwegian Sea. Rapports et Procès-Verbaux des Réunions du Conseil International pour l'Exploration de la Mer, 188: 49–58.

Dethlef, D. 1995. Die Laptevesee – eine Schlüsselregion für den Fremdstoffeintrag in das arktische Meereis. Dissertation (in German), Mathematisch-Naturwissenschaftliche Fakultät der Christian-Albrechts-Universität zu Kiel.

Gjevik, B., Nost, E., and Straume, T. 1994. Model simulations of the tides in the Barents Sea. Journal of Geophysical Research, 99: 3219–3230.

Harms, I. H. 1992. A numerical study of the barotropic circulation in the Barents and Kara Seas. Continental Shelf Research, 12: 1043–1058.

Harms, I. H. 1994. Numerische Modellstudie zur winterlichen Wassermassenformation in der Barentssee. Dissertation (in German), Berichte aus dem Zentrum für Meeres- und Klimaforschung, Reihe B: Ozeanographie, Nr. 7.

Harms, I. H. In press. Modelling the dispersion of ^{137}Cs and ^{239}Pu released from dumped waste in the Kara Sea. Journal of Marine Systems.

Hibler, W. D. III 1979. A dynamic thermodynamic sea ice model. Journal of Physical Oceanography, 9: 815–846.

Kowalik, Z., and Proshutinsky, A. Y. 1993. Diurnal Tides in the Arctic Ocean. Journal of Geophysical Research, 98: 16 449–16 468.

Kowalik, Z., and Proshutinsky, A. Y. 1994. The Arctic Ocean Tides. *In* The Polar Oceans and Their Role in Shaping the Global Environment. The Nansen Centennial Volume. pp. 137–158. Ed. by O. M. Johannessen, R. D. Muench and J. E. Overland. Geophysical Monography 85, AGU, Washington DC, USA.

Lemke, P., Owens, W. B., and Hibler, W. D. 1990. A coupled sea ice mixed layer pycnocline model for the Weddell Sea. Journal of Geophysical Research, 95: 9513–9525.

Levitus, S. 1982. Climatological atlas of the world ocean. NOAA Professional Paper 13, U.S. Department of Commerce: National Oceanic and Atmospheric Administration, Boulder, Colorado, USA.

Loeng, H. 1991. Features of the physical oceanographic conditions of the Barents Sea. Polar Research, 10: pp. 5–18.

Loeng, H., Ozhigin, V., and Ådlandsvik, B. 1995. Water fluxes through the Barents Sea. ICES Mini Symposium on Arctic Processes, Aalborg, Denmark, ICES CM 1995/Mini: 10.

Martin, S., and Cavalieri, D. J. 1989. Contribution of the Siberian shelf Polynya to the Arctic Intermediate and Deep Water. Journal of Geophysical Research, 94: 12725–12738.

Maykut, G. A., and Untersteiner, N. 1971. Some results from a time dependent thermodynamic model of sea ice. Journal of Geophysical Research, 76: 1550–1575.

Maykut, G. A. 1986. The surface heat and mass balance. *In* Geophysics of sea ice, pp. 395–463. Ed. by N. Untersteiner. NATO ASI Series B, Physics. Vol. 146 Plenum Press, New York.

Midttun, L. 1985. Formation of dense bottom water in the Barents Sea. Deep-Sea Research, 32: 1233–1241.

Nansen, F. 1906. Northern Waters: Captain Roald Amundsen's oceanographic observations in the Arctic Seas in 1901, with a discussion of the origin of the bottom waters of the Northern Seas. Vid.-selskab Skrifter I, Mat.-Naturv. Klasse 1906, No. 3, Jacob Dybwads Forlag, Christiania, 145 pp.

Norwegian Polar Research Institute. 1990. Den Norske Los/ Arctic Pilot. Sailing Direktions Vol. 7, 2nd edn. The Norwegian Hydrographic Service, Stavanger.

Parkinson, C. L., and Washington, W. M. 1979. A large scale numerical model of sea ice. Journal of Geophysical Research, 84: 311–337.

Quadfasel, D., Rudels, B., and Kurz, K. 1988. Outflow of dense water from a Svalbard fjord into the Fram Strait. Deep-Sea Research, 35: 1143–1150.

Rudels, B., and Quadfasel, D. 1991. Convection and deep water formation in the Arctic Ocean–Greenland Sea System. Journal of Marine Systems, 2: 435–450.

Schauer, U. 1995. The release of brine enriched shelf water from Storfjord to the Norwegian Sea. Journal of Geophysical Research, 100: 16 015–16 028.

Semtner, A. J. 1976. A model for the thermodynamic growth of sea ice in numerical investigations of climate. Journal of Physical Oceanography, 6: 379–389.

Smith, S. D., Muench, R. D., and Pease, C. H. 1990. Polynias and Leads: an overview of physical processes and environment. Journal of Geophysical Research, 95: 9461–9479.

Stronach, J. A., Backhaus, J. O., and Murty, T. S. 1993. An update on the numerical simulation of oceanographic processes in the waters between Vancouver Island and the mainland: the GF8 model. Oceanography and Marine Biology Annual Review, No. 31, pp. 1–86.