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The effects of the water flow through the Canadian Archipelago in a global ice–ocean model

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Abstract. Numerical experiments are conducted with a global ice–ocean model in order to evaluate the influence of the water flow from the Arctic Ocean to Baffin Bay through the Canadian Archipelago on the water-mass properties of the Arctic Ocean and adjacent seas and, more generally, on the global ocean circulation. The results indicate that this flow plays a significant role in controlling the freshwater budget of the Arctic Ocean. When the Canadian Archipelago passage is open in the model, the Arctic pycnocline experiences a noticeable increase in salinity. Furthermore, the flow of relatively fresh Arctic waters through the passage yields a pronounced decrease of surface salinity and density in the Labrador Sea, which leads to a diminution of convective activity there. As a result, the North Atlantic Deep Water outflow in the model is reduced by about 5%. Deep convection in the Norwegian Sea exhibits almost no change, and this despite a weakening of the inflow of relatively fresh Arctic waters through Fram Strait.

Introduction

The Canadian Archipelago is a large and complex system of less than 250-m-deep channels through which cold and relatively fresh upper waters from the Arctic Ocean enter Baffin Bay at an annual mean rate of 1 to 2 Sv (1 Sv = 10^6 m^3 s^{-1}) [*Rudels*, 1986]. The freshwater flux associated with this outflow seems to take an important part in determining the freshwater balance of the Arctic Ocean. Several studies [e.g., *Aagaard and Carmack*, 1989; *Steele et al.*, 1996] suggest that this freshwater flux is larger than the one associated with the water export through Fram Strait, the other major exit door for Arctic surface waters. In fact, the volume transports in the upper 200 m are approximately the same but, as the salinity of waters flowing through the Canadian Archipelago is lower, the contribution of the archipelago throughflow to the Arctic freshwater budget is higher [*Steele et al.*, 1996].

The water flow through the Canadian Archipelago acts as a source of buoyancy for the upper layers of Baffin Bay and of the Labrador Sea, and thus tends to weaken the deep convection taking place in the Labrador Sea. Since this convection, together with that occurring in the Greenland-Iceland-Norwegian (GIN) seas, is at the origin of the formation of North Atlantic Deep Water (NADW) [e.g., *Clarke and Gascard*, 1983], it is expected that the Canadian Archipelago throughflow affects the global thermohaline circulation.

Global ice-ocean models used for climate studies have generally coarse grids because of the heavy computational cost of the model and the need to carry out relatively long integrations to reach a quasi-equilibrium state of the thermohaline circulation. Such grids do not allow to have a good representation of the complicated bathymetry in the archipelago area and, as a consequence, the passage through the Canadian Archipelago is generally considered to be closed in these models. To enlighten the consequences of this assumption, we examine the effects of the water flow through the Canadian Archipelago in a coarse-resolution, global ice–ocean model. The detailed structure of the flow is obviously not simulated. In particular, it is impossible with such a model to represent the various channels through the archipelago. The goal of this study is only to estimate the impact of a direct communication between the Arctic Ocean and Baffin Bay on the water-mass properties and circulation of the Arctic Ocean and peripheral seas, and on the global ocean circulation.

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The Model, Forcing, and Experimental Design

The model used here results from the coupling of a primitive-equation, free-surface ocean general circulation model [*Deleersnijder and Campin*, 1995] with a comprehensive thermodynamic–dynamic sea-ice model [*Fichefet and Morales Maqueda*, 1997]. The model equations are solved on *Arakawa*'s [1966] B-grid. To cope with the singularity at the North Pole associated with the geographical spherical coordinates, two spherical grids connected in the equatorial Atlantic are employed: a grid with its poles located on the geographical equator for the North Atlantic and the Arctic, and a classical latitude–longitude grid for the rest of the World Ocean [*Deleersnijder et al.*, 1993]. The horizontal resolution is of 3° by 3°, and there are 20 vertical levels in the ocean. The water flow through Bering Strait is parameterized as a linear function of the cross-strait sea-level difference in accordance with the geostrophic control theory [*Toulany and Garrett*, 1984; *Goosse et al.*, 1997].

The model is driven by surface fluxes of heat, freshwater, and momentum determined from atmospheric data by using bulk formulas. Inputs fields consist of annual mean river-runoff data and of monthly climatological surface air temperatures and humidities, precipitation rates, cloud fractions, surface winds, and wind stresses. In addition, a relaxation towards observed annual mean salinities [*Levitus*, 1982] is applied in the 10-m-thick surface grid box with a time constant of two months in order to prevent any salinity drift. For more details about the forcing, see Goosse et al. [1997].

Two quasi-equilibrium experiments of 750-yr duration are performed with the model. In the first one (CL), no water transport is allowed between the Arctic Ocean and Baffin Bay. In the second one (OP), a two-grid-point-wide channel is opened west of Greenland, the minimum width to have a transport on a B-grid. This channel will be hereafter referred to as the Arctic–Baffin Bay passage (ABBP). Its depth is taken as 150 m, i.e., the model shelf depth in that region. No ice transport is permitted through ABBP. This is a reasonable approximation since this transport is known to be very small [e.g., *Aagaard and Carmack*, 1989], the ice present in the region of the Canadian Archipelago being generally fastened to the coast (landfast ice) [*Flato and Brown*, 1996].

Results and Discussion

The modelled volume transport through ABBP amounts to about 1 Sv on an annual average, which is within the range of observational estimates and close to the values obtained with models using a finer grid [e.g., *Holland et al.*, 1996]. The effects of this flow are particularly pronounced in Baffin Bay. In experiment CL (ABBP closed), the relatively cold and fresh waters located in the top 70 m of the bay are advected southwards by the wind. Below this layer, there is an intrusion of waters originating from the Atlantic that gives rise to a maximum of both temperature and salinity at depth. When ABBP is open (experiment OP), the flow is southwards in the first 200 m, and it carries colder and fresher waters from the north. As a consequence, the water is below 0°C from the surface to a depth of 250 m in Baffin Bay, the Atlantic waters being present only at greater depths, in better agreement with observations (Figure 1).

Once outside Baffin Bay, the waters coming from the Arctic Ocean are advected southwards along the eastern coast of Canada by the Labrador Current. The annual mean volume transport associated with this current is increased by approximately 1.5 Sv in experiment OP. (This strengthening is larger than the magnitude of the ABBP throughflow because of recirculation in the Labrador Sea.) The Arctic waters are then transported eastwards by the North Atlantic subpolar gyre, whose strength is enhanced by about 1 Sv. This drift can be

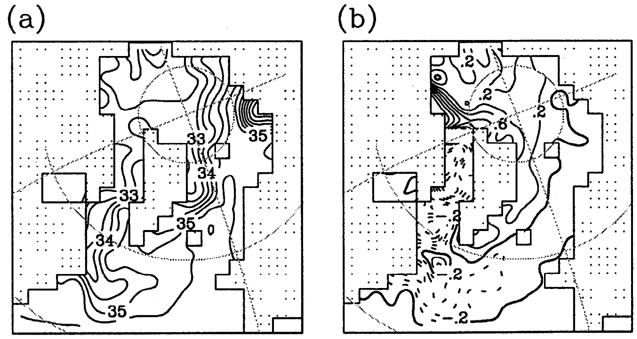

Figure 2. (a) Geographical distribution of the annual mean salinity at a depth of 120 m in the Arctic Ocean and peripheral seas for experiment OP. (b) Same as Figure 2a, except for the difference of salinity between experiments OP and CL. Units are psu; contour intervals are 0.5 psu (a) and 0.1 psu (b). Negative values are dashed.

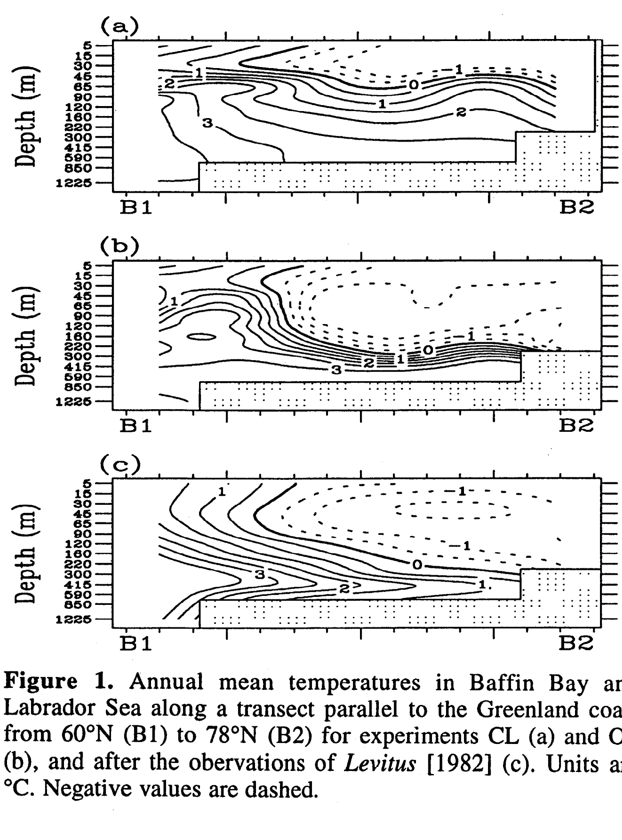

Figure 1. Annual mean temperatures in Baffin Bay and Labrador Sea along a transect parallel to the Greenland coast from 60°N (B1) to 78°N (B2) for experiments CL (a) and OP (b), and after the obervations of *Levitus* [1982] (c). Units are °C. Negative values are dashed.

deduced from Figure 2a, which shows the annual mean salinities at a depth of 120 m in the Arctic Ocean and peripheral seas for experiment OP. Figure 2b, which gives the differences in annual mean salinity at a depth of 120 m between the two experiments, indicates that opening ABBP leads to a freshening that reaches 1.8 psu in the northern part of Baffin Bay and that decreases progressively along the path of the flow to less than 0.1 psu in the Norwegian Sea. This freshening generally reduces the difference with Levitus (1982) observations. For example, the magnitude of the error in the Labrador Sea is smaller by about 0.1 psu in experiment OP, the model being a little too fresh in experiment OP and too salty in experiment CL .

The decrease of surface salinity in the Labrador Sea is responsible for a noticeable weakening of the deep convection that occurs in this sea. Furthermore, the area where intense convective mixing takes place is reduced and is moved southwards. The spot of higher salinities in the central Labrador Sea in experiment OP as compared to experiment CL (see Figure 2b) is a consequence of this shift. Convection does not arise there in experiment CL but does in experiment OP, thus bringing saltier water from the deep ocean towards the surface. In accordance with the changes in convection, the annual mean meridional overturning is reduced by 2.5 Sv at the latitudes of the Labrador Sea (Figure 3). As a result, the amount of NADW exiting the Atlantic Ocean at 30°S is decreased by 0.9 Sv (5%). No significant modification of the water-mass properties of the Atlantic Ocean is noticed; the annual mean meridional heat transport at 20°N in this ocean is reduced by 4% (i.e., 0.04×10^{15} W).

The surface salinity averaged over the GIN seas exhibits a slight increase in experiment OP. This is due to a lower inflow of freshwater from the Arctic Ocean, which is caused by a smaller southward volume transport through Fram Strait and higher salinities in the upper Arctic Ocean (see Figure 2 and below). Actually, the salinity increase in the GIN seas is restricted to the western Greenland Sea, a highly stratified area where deep convection does not occur in the model nor in the real ocean. In the regions of strong convective mixing (i.e., the eastern Greenland Sea and the Norwegian Sea), the salinities remain more or less the same, so that the convective activity and the overturning north of 65°N are nearly unchanged.

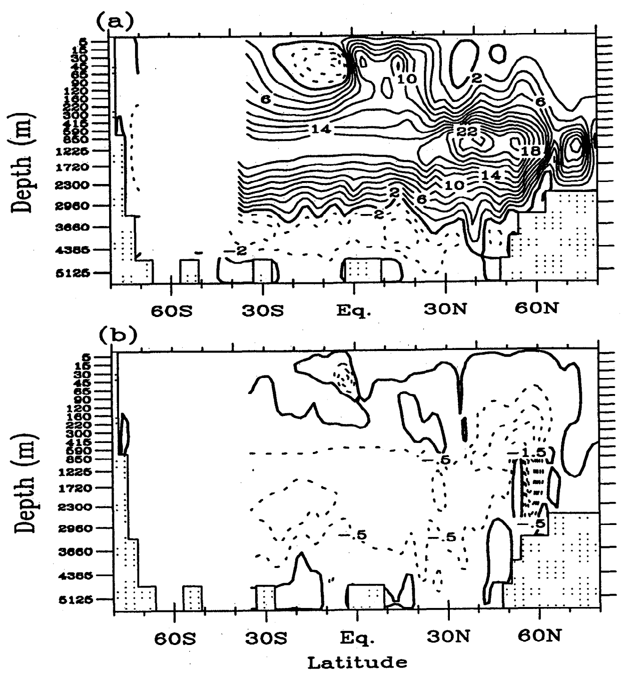

Figure 3. (a) Contours of the annual mean meridional overturning streamfunction in the Atlantic basin for experiment OP. (b) Same as Figure 3a, except for the difference of streamfunction between experiments OP and CL. Flow is clockwise around solid contours. Units are Sv; contours intervals are 2 Sv (a) and 0.5 Sv (b).

The Arctic pycnocline is appreciably saltier in experiment OP (the salinity increase averages 0.3 psu at a depth of 120 m). Hence, the structure of the pycnocline is in better agreement with that of observed profiles, altough the simulated salinities are still too low (0.7 psu lower than Levitus' data on average at 120 m). The differences of salinity between the two runs are particularly large over the continental shelf located north of Canada (see Figure 2). In this region, the surface net freshwater flux is positive (i.e., the ocean gains freshwater from the surface). As the water exchanges between the shelf and the open ocean are rather limited in experiment CL, the incoming freshwater remains over the shelf, thus leading to salinities there that are up to 2 psu lower than in the observations. In experiment OP, part of the surface freshwater flux is exported by the flow through the Canadian Archipelago. Furthermore, this transport induces an inflow of saltier waters from the open-ocean pycnocline onto the shelf. These two processes combine to produce higher and much more realistic salinities in this sector.

The salinity increase in the Arctic pycnocline is a consequence of a reorganization of the freshwater balance of the Arctic Ocean. When no water transport is allowed through the Canadian Archipelago, the surface freshwater fluxes and the freshwater imports from the North Pacific and the North Atlantic (the sum of these three fluxes will be hereafter referred to as F_{INP}) must be balanced by the freshwater export through Fram Strait. Thus, one has

$$\frac{S_{FRAM}^{CL} - S_{REF}}{S_{REF}} V_{FRAM}^{CL} = F_{INP} \tag{1}$$

where V_{FRAM}^{CL} is the volume export through Fram Strait, S_{FRAM}^{CL} is the salinity of the waters exiting the Arctic Ocean through Fram Strait, and S_{REF} is a reference salinity taken as 34.8 psu following *Aagaard and Carmack* [1989]. Volume conservation implies that V_{FRAM}^{CL} must be equal to V_{INFLOW}, the total volume transport into the Arctic Ocean. Assuming that V_{INFLOW} and F_{INP} do not change when ABBP is open, the freshwater budget of the Arctic Ocean for experiment OP may be written as

$$\frac{S_{FRAM}^{OP} - S_{REF}}{S_{REF}} V_{FRAM}^{OP} + \frac{S_{ABBP} - S_{REF}}{S_{REF}} V_{ABBP} = F_{INP} \tag{2}$$

where S_{FRAM}^{OP} (V_{FRAM}^{OP}) and S_{ABBP} (V_{ABBP}) are the salinity (volume export) in Fram Strait when ABBP is open and in ABBP, respectively. Again, the conservation of volume yields

$$V_{FRAM}^{OP} + V_{ABBP} = V_{INFLOW} \tag{3}$$

which signifies that the total inflow is now balanced by the export through Fram Strait and the ABBP rather than through Fram Strait alone as in experiment CL.

By combining (1), (2), and (3), one obtains

$$S_{FRAM}^{OP} = S_{FRAM}^{CL} + \frac{V_{ABBP}}{V_{INFLOW}} \left(S_{FRAM}^{OP} - S_{ABBP} \right) \tag{4}$$

As S_{FRAM}^{OP} is higher than S_{ABBP}, S_{FRAM}^{OP} must be greater than S_{FRAM}^{CL}. Those equations mean that the freshwater flux associated with the flow through ABBP in experiment OP is higher than the one associated with the same amount of fluid which was exported through Fram Strait in experiment CL, the salinity in Fram Strait being higher than in ABBP. To balance the Arctic freshwater budget in experiment OP, the freshwater export must decrease somewhere compared to experiment CL, all the imports remaining nearly constant. This is achieved by an increase of the salinity of the outflow through Fram Strait in experiment OP. Since the flow through Fram Strait is fed by waters originating from various parts of the Arctic basin, the salinity increase tends to be generalized to the whole Arctic area (see Figure 2). By introducing into (4) the values of V_{INFLOW}, V_{ABBP}, S_{FRAM}^{OP}, and S_{ABBP} generated by the model (i.e., 3.0 Sv, 1.0 Sv, 33.67 psu, and 31.80 psu, respectively), one gets a salinity difference of 0.45 psu, which compares favourably with the salinity difference of 0.31 psu produced by the model.

The above calculation is based on the assumption that the surface freshwater flux F_{INP} and the volume import V_{INFLOW} remain the same when ABBP is open. This is confirmed by the model for V_{INFLOW} (a change smaller than 0.1 SV is observed) as well as for the freshwater inputs from the North Pacific and the North Atlantic (the differences are less than 2%). However, the surface freshwater flux induced by the relaxation towards climatological salinities is somewhat modified, which explains why the correspondence between the salinity difference calculated from (4) and the model results is not perfect. It is worth stressing that this restoring is necessary to avoid dramatic errors in the modelled salinity field, errors that are due to uncertainties in the freshwater forcing and to model shortcomings. For instance, *Weatherly and Walsh* [1996] did not use any restoring in their regional model of the Arctic Ocean and adjacent seas. After less than 30 years of simulation, the computed surface salinities in the Norwegian Sea were about 10 psu lower than in the observations. Such a drift in our global model would shut down deep-water formation in the Norwegian Sea and, consequently, would lead to an unrealistic World Ocean's thermohaline circulation. The restoring freshwater flux is 0.016 Sv lower in experiment CL as compared to experiment OP over the Arctic basin and higher by roughly the same amount over Baffin Bay and the Labrador Sea. This means that, with ABBP closed, the relaxation gives rise to a freshwater transport from the Arctic Ocean to Baffin Bay of approximately 0.02 Sv compared to the case with ABBP open. This unphysical transport is notably weaker than the simulated freshwater transport through the Canadian Archipelago, which amounts to 0.07 Sv on an annual average (this value is somewhat higher than current estimates (0.03 SV after *Aagaard and Carmack* [1989]; 0.04 Sv after *Steele et al.* [1996]) as the model tends to underestimate the salinity of the Arctic Ocean). These results suggest that the change in the relaxation forcing does not play a dominant role in the model response. Nevertheless, it must be borne in mind that this change has a non-zero effect when interpreting the results of the sensitivity experiments presented here.

Summary and Conclusions

The main goal of this study was to evaluate the effects of the water flow through the Canadian Archipelago in a global ice–ocean model suitable for climate studies. When the Arctic Ocean and Baffin Bay are allowed to exchange water in the model, the Arctic pycnocline becomes significantly saltier. This demonstrates that the outflows of water from the Arctic Ocean are important for the characteristics of the Arctic pycnocline waters, a balance between the inflows and the outflows controlling their salinity and temperature.

Many ice–ocean models have a tendency to underestimate the strength of the Arctic pycnocline. This bias is very often attributed to an excessive mixing of the fresh surface layer with the more saline Atlantic layer. In principle, the incorporation of comprehensive turbulence-closure schemes into the models should reduce the error [e.g., *Weatherly and Walsh*, 1996]. However, the problem is not restricted to vertical mixing. We have shown that a better ventilation of the Arctic Ocean thanks to the opening of ABBP leads to a more realistic simulation of the Arctic pycnocline.

The Canadian Archipelago throughflow induces in our model a pronounced freshening of the Labrador Sea. The immediate effect of this is an appreciable weakening of the convective activity in this sea. As a result, the Atlantic thermohaline overturning experiences a decrease in strength of about 5% at 30°S. The weakening of the convective mixing in the Labrador Sea is not compensated by an intensification of convection in the GIN seas. The freshwater import into these seas through Fram Strait is smaller when ABBP is open. However, the resulting positive salinity anomaly is restricted to the western Greenland Sea. In the eastern Greenland Sea and in the Norwegian Sea, where deep convection occurs in the model, the changes in salinity are insignificant.

We have carried out a complementary experiment in which the width of the modelled strait has been reduced by a factor five by decreasing locally the size of the grid-mesh, this modification allowing a passage width in better agreement with the real bathymetry. In this case, the volume transport from the Arctic Ocean to the Baffin Bay amounts to 0.5 Sv. An analysis of the results leads to the same kind of conclusion than the one obtained from experiments OP and CL. For example, because of this 0.5 Sv flow, the salinities in the Arctic pycnocline at 120 m are 0.2 psu higher and the NADW outflow is 3% lower than in an experiment with ABBP closed.

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