Feedback mechanisms affecting the thermohaline circulation with atmosphere and sea ice

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Introduction

Changes in high latitude surface salinity have a strong effect on the North Atlantic Deep Water Formation (NADWF) which appears to be very important in driving the global thermohaline conveyor belt. Natural variations of sea surface salinity and sea ice have been observed in the North Atlantic, namely the Great Salt Anomaly (GSA) of the late sixties and seventies.
When dealing with climate variability one must consider the sensitivity of the climate system to perturbations.

In order to include the atmospheric heat transport mechanisms we coupled an atmosphere energy balance model with a 3-D ocean general circulation model which includes a thermodynamic sea ice model.

We explore the feedback mechanisms in the ocean-atmosphere-sea ice system affecting the thermohaline circulation (THC) under perturbations in sea surface salinity at high latitudes.

Model

The atmosphere energy balance model consists of vertically integrated equations for mixing ratio of water vapour q_v and atmosphere temperature T :

$$L_v \, \partial_t q_v \; + \; \nabla \cdot (L_v \, v \, q_v) \; = \; L_v \, (E \; - \; P) \qquad (1)$$

$$C_p \, \partial_t T \; + \; \nabla \cdot (C_p \, v \, T) \; = \; Q_R + Q_L + Q_S \qquad (2)$$

Q_R : Radiative flux, $Q_L = L_v \, (E \; - \; P)$: Latent heat release, Q_S : Sensible heat flux at the air-sea interface. Combining equations (1,2) we get the vertically integrated energy equation:

$$C_p \, \partial_t T \; + \; \nabla \cdot (C_p v T) \; + \; \nabla \cdot (L_v v q) \; = \; Q_R^{top} \; - \; F_{oa}$$

$F_{oa} = Q_R^{bot} - L_v E - Q_S$: Ocean-atmosphere heat flux, calculated by simplified bulk formulas. At high latitudes the **transient eddies** dominate the meridional heat transport. The mean transports are held fixed while the transient eddies are paramiterised based on baroclinic instability theory (Green, 1970).

$$\overline{v'T'} \; = \; - \; K_s \; \left(\frac{\partial \, T}{\partial y}\right)^2$$

$$\overline{v'q'} \; = \; rh \; \frac{\partial \, q_s}{\partial T} \; \overline{v'T'} \qquad ,$$

rh : relative humidity, q_s : saturation mixing ratio.
The sensible and latent eddy heat transports are calculated in terms of the surface quantities and are tuned to reproduce the current climate. The planetary albedo is parameterized in terms of the surface air temperature. No seasonal cycle is considered.

The ocean model is the GFDL primitive equation model (Pacanowski et al.,1993,1991). The model domain is a 70^0 wide sector from 66^0 S to 80^0 N. It includes an idealized Greenland-Scottland ridge.

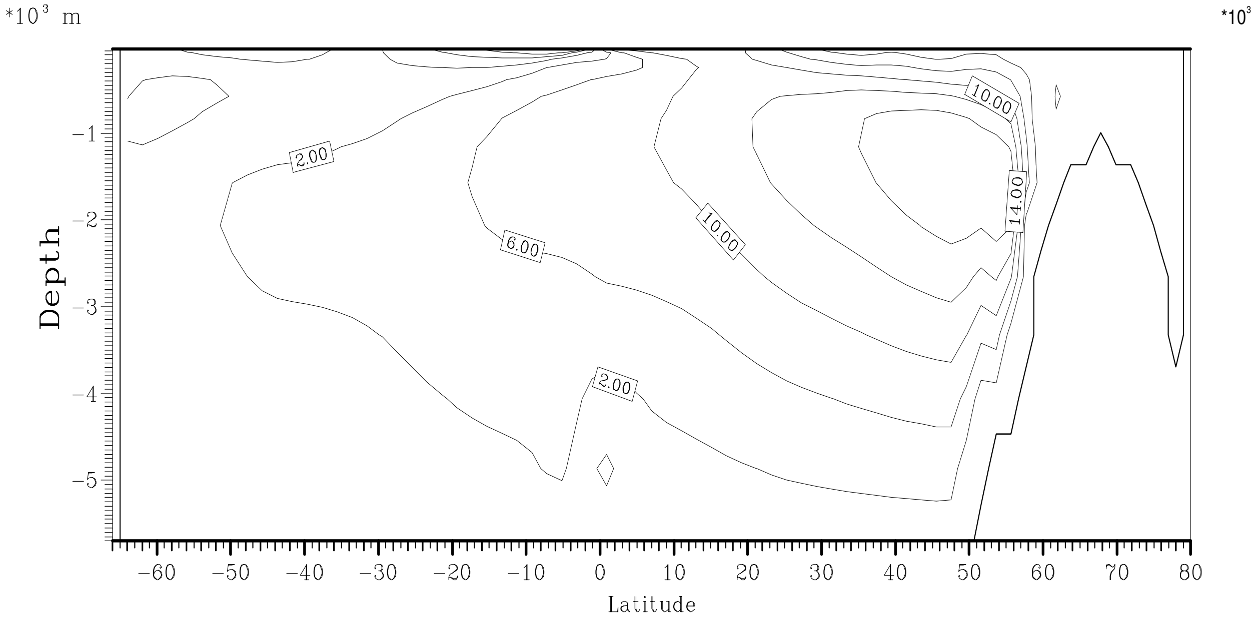

Figure 1: Stream function for the zonally integrated transport at the end of the spin up in Sv ($10^6 \frac{m^3}{s}$) and initial state for subsequent experiments. The maximal overturning rate is about 22 Sv. The NADWF takes place between 50^0 N and 60^0 N while sea ice is found north of 65^0 N.

Experiments and Discussion

We start from a steady state circulation shown in figure 1. Smaller pertubations compared to the GSA induce a breakdown of the THC under mixed boundary conditions (fixed atmosphere temperature and surface fresh water flux). The main deficiency of that b.c. is that the atmosphere is held fixed although the air-sea heat flux is strongly reduced.
In our coupled system the THC does not collapse (figure 3) with a perturbation in salinity of 0.5 psu between $50^0 - 70^0$ N, which is much stronger than the GSA.

The response of the coupled system (shown in figures 3,4,5,6) to that perturbation is a combination of different mechanisms:

- Sea ice cools the overlying air by changing the albedo and by hindering the ocean-atmosphere exchange.
- The northward atmospheric heat transport is strengthend and the air at high latitudes is warmed. So the cooling of the air by sea ice is partly compensated by a stronger eddy activity (figures 6,4). If the meridional heat transport is constant the atmospheric temperature is only effected by a local heat balance. Then a larger sea ice extent will reduce the poleward heat and salt transports and the THC will break down.
- Through a higher cyclonic activity the fresh water flux $P-E$ changes (figure 5). However, this change turns out to be of minor importance for the THC. In the model precipitation is not transformed into snow.

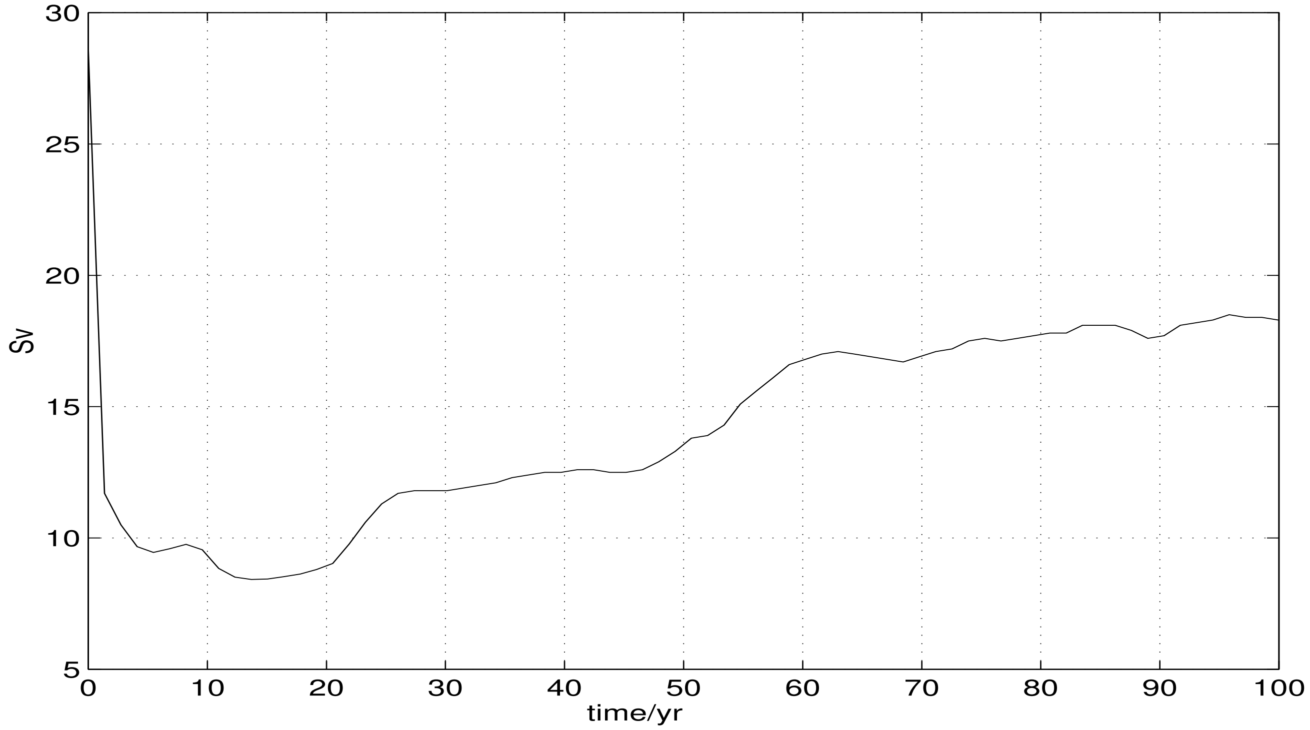

Figure 2: Timeseries of the maximal overturning stream function in the northern hemisphere in Sv. The THC in the coupled system is stabilized and does not break down.

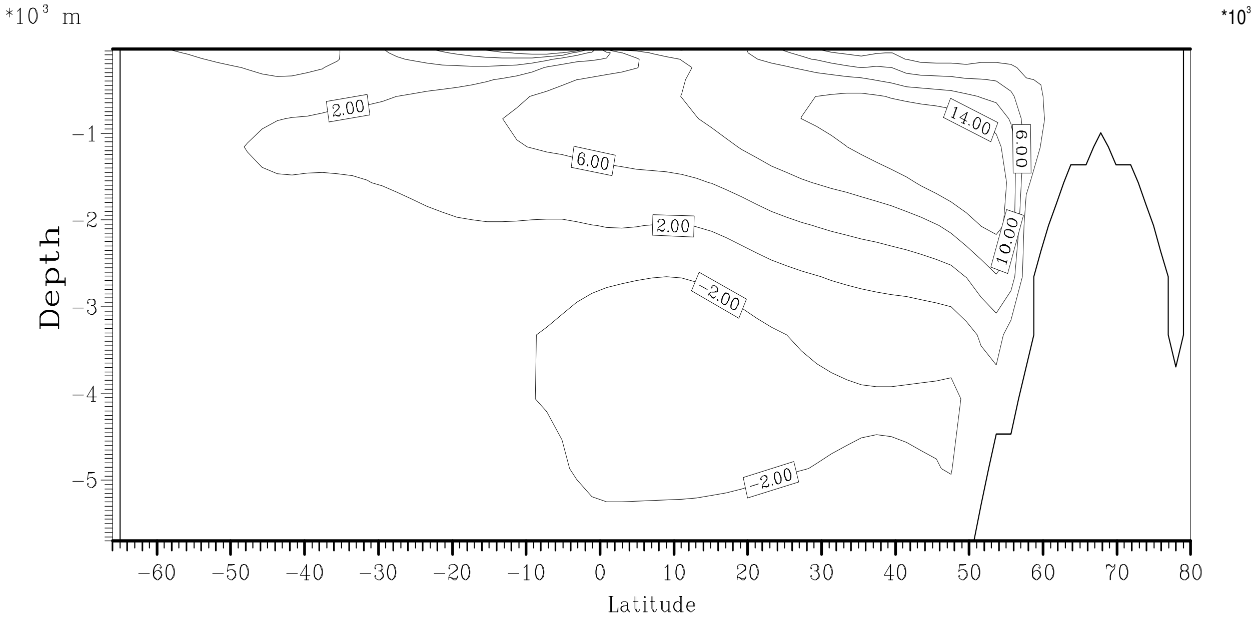

Figure 3: Stream function in the coupled system 110 yrs after the perturbation at high latitudes. The maximal overturnig rate is about 18 Sv.

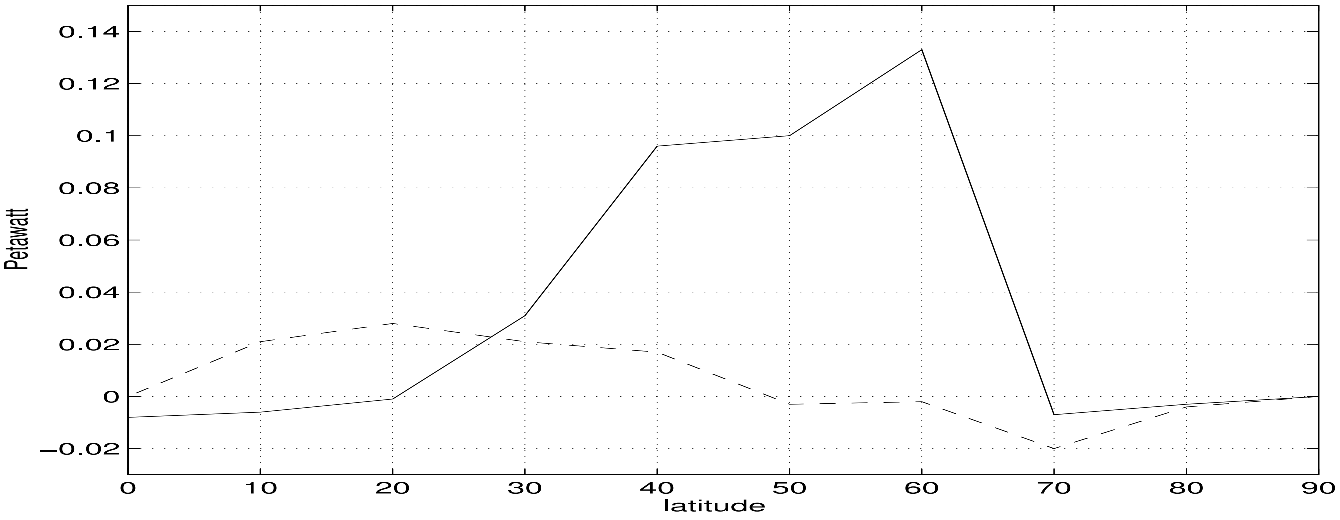

Figure 4: Change of the latent (− −) and sensible (−) heat transports in the northern hemisphere, 110 yrs after the perturbation.

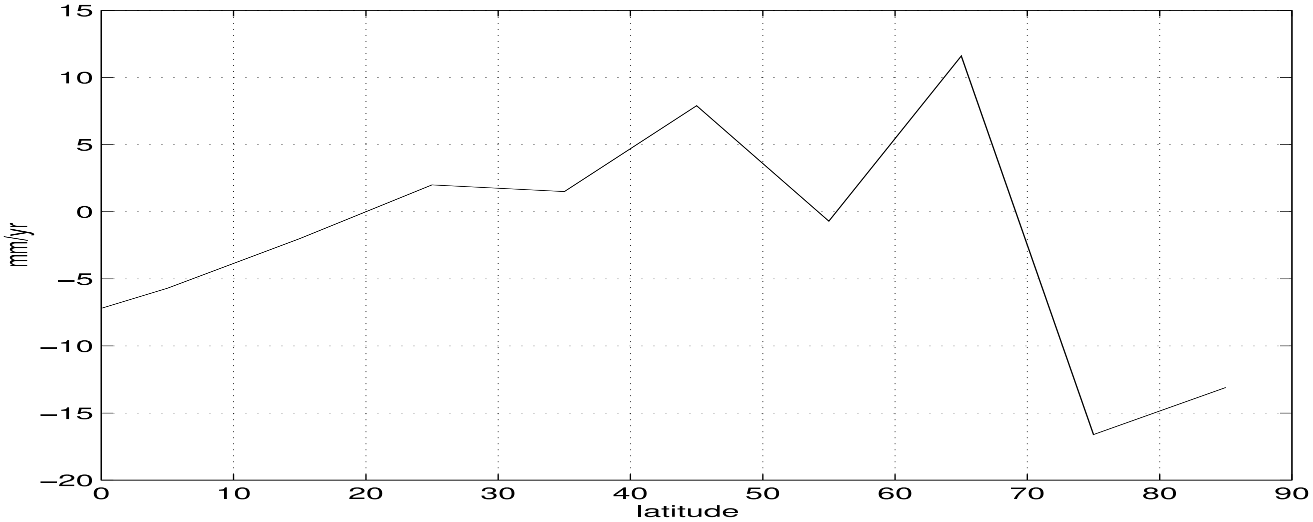

Figure 5: Change of $P - E$ in the northern hemisphere after 110 yrs.

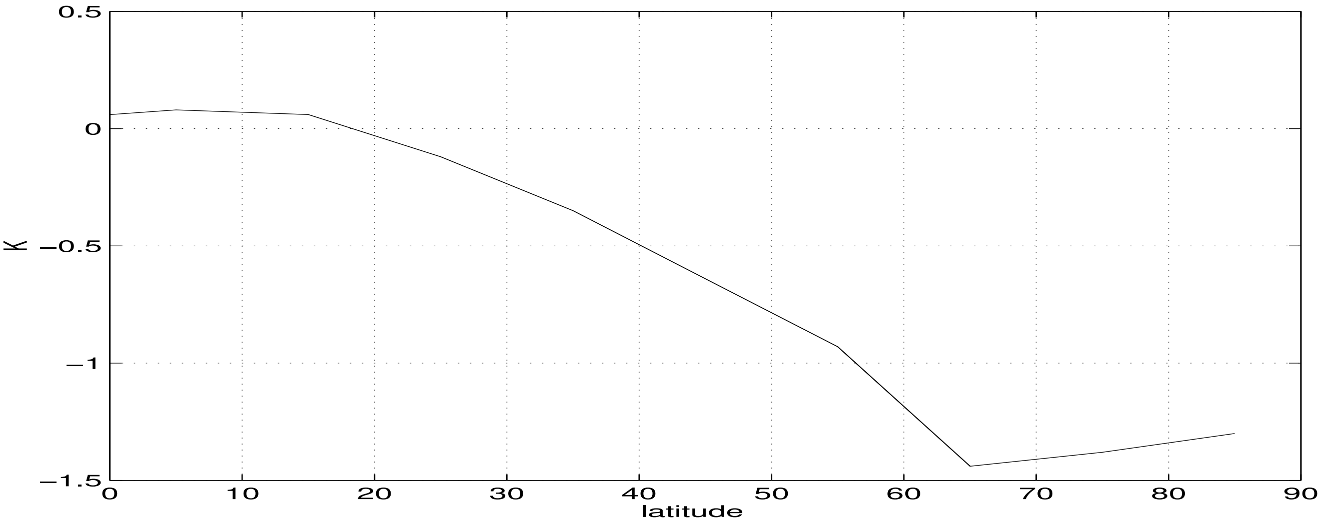

Figure 6: Change of the atmospheric surface temperature in the northern hemisphere 110 yrs after the perturbation at high latitudes.

Conclusions

Several investigators have observed oscillations of the THC on decadal time scale, (Weaver and Sarachick,1991, Yang and Neelin,1993, Zhang et al.,1994). However, none of them included an interactive atmosphere model, so their results should be questioned in terms of the coupled system.

The stability of the THC and NADWF depends on different feedback mechanisms in the coupled ice-ocean-atmosphere system. Some of them, like the atmospheric heat transport, which are deemed to be important for the THC, are included in our model.

It is concluded that the atmospheric heat transport has a stabilising effect on the circulation in the coupled model.

A logical next step is to apply a more complex model to investigate the feedbacks identified in our model.