The Influence of a near-bottom Transport Parameterization on the Sensitivity of the Thermohaline Circulation

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Abstract

The effect of a near bottom transport scheme on the sensitivity of the thermohaline circulation is analyzed in a coupled model. In this model with idealized geometry of the Atlantic, it is shown that in the presence of a northern source of deep water, an accurate representation of the overflow process has a stabilizing effect on the thermohaline circulation for subpolar sea surface salinity perturbations. The large scale overturning circulation can be maintained in the presence of a continued deep water formation north of the sill. Experiments suggest that without a sufficient coupling across a sill in the northern North Atlantic, the response of the ocean's circulation to subpolar atmospheric variability may be too strong.

1 Introduction

The northern hemisphere oceanic heat transport is mostly due to the large-scale thermohaline circulation (THC). One of the THC's main driving force is the sinking of dense water masses at high latitudes. Part of this water is formed at intermediate depths and slides down the continental slopes. In the northern North Atlantic, this water contributes to the deep water renewal (Aagaard et al., 1985). Experiments with ocean circulation models show a strong dependence of the THC on the presence of the North Atlantic overflow water masses (Döscher et al., 1994; Gerdes and Köberle, 1995; Roberts and Wood, 1997; Döscher and Redler, 1997; Redler and Böning, 1997). The North Atlantic is thus a region of particular importance because changes in the rate of deep water production can have a profound effect on climate.

In nature, several water masses contribute to the North Atlantic Deep Water, viz Labrador Sea Water, overflow water of intermediate waters from the Nordic Seas, and entrained salty subtropical water (Dickson and Brown, 1994). About 5.6 Sv overflows the Greenland-Scotland Ridge on either side of Iceland at $450 - 850$ m in depth for the Holocene. Paleoclimatic studies (e.g. Imbrie et al., 1992) suggest that if one of these sources is shut off, the THC switches into a different mode. Because dense water on the continental slopes can contribute to the deep water renewal, the representation of this water mass is potentially important for climate sensitivity.

The explicit simulation of overflow water is very critical because these processes depend on details of topography (see e.g. Roberts and Wood, 1997; Redler and Böning, 1997) and model formulation. Winton et al. (1997) have compared the down-slope flow of dense water in a z-coordinate and an isopycnal coordinate model. While the isopycnal coordinate model moves water down-slope without entrainment, the z-coordinate model has very strong numerical entrainment and dense signals do not reach the bottom. In z-coordinate models, down-slope transport is treated as alternating advection and convection thereby loosing bottom water mass characteristics (Gerdes, 1993 a). The two models of Winton et al. (1997) converge when the z-coordinate model resolves the bottom layer horizontally and vertically. Such a resolution is, however, prohibitive for climate sensitivity studies.

It is therefore desirable to parameterize explicitly the near bottom transport over sloping topography in coarse resolution models. Because the drawbacks of other model concepts (horizontal pressure gradient error in terrain-following, vertical resolution problems in isopycnic models), Beckmann and Döscher (1997) formulate a bottom boundary layer (BBL) model combining the virtues of the sigma-transformation with the usual z-coordinate model. Their basic idea is that the tracer points in the bottom boundary layer model and the z-coordinate model are identical and therefore, concentrations can be exchanged directly. They formulate a conditional diffusive BBL model to represent slope convection in the case of dense water above light water. Their model parameterizes advective processes near the bottom by a large mixing coefficient if the condition for down-slope convection is satisfied.

Another representation of the BBL has been proposed by Gnanadesikan et al. (1998). His model includes a thin BBL in order to resolve friction an explicitly calculated pressure gradient term within this terrain-following ($\sigma-$) layer. Calculating horizontal pressure gradient terms in $\sigma-$coordinate models have, however, the possibility of large errors for coarse resolution models with steep topography (Haney, 1991). Killworth and Edwards (1998) develop the BBL model approach further by introducing a spatial and temporal variable benthic boundary layer. Their dynamics for the benthic boundary layer are based on boundary layer physics. Killworth and Edwards' (1998), Gnanadesikan et al.'s (1998) and Beckmann and Döscher's (1997) experiments with idealized continental slopes show the dense water transport over topography, when a BBL parameterization scheme is included.

Here, the sensitivity of the THC is analyzed in an coupled atmosphere-ocean-sea ice model (Lohmann and Gerdes, 1998) with idealized geometry of the Atlantic ocean. In the model, a slope convection scheme is included, following the concept of Beckmann and Döscher (1997). We examine how a bottom boundary layer included in the model can affect the response to a subpolar salinity perturbation.

Sensitivity experiments with ocean only (Bryan, 1986; Mikolajewicz and Maier-Reimer, 1994) and fully coupled models (Manabe and Stouffer, 1995; Schiller et al., 1997) show that the THC is largely affected by surface haline perturbations interrupting deep water formation. Those numerical experiments with the present day circulation as the reference climate have been viewed as paradigms for cooling events observed in the paleoclimatic record (Fairbanks 1989; Keigwin et al., 1991; Dansgaard et al., 1993). A more recent freshening event which interrupted temporarily deep water renewal in the Labrador Sea has been observed during the Great Salinity Anomaly in the late sixties (Dickson et al., 1988; Lazier, 1980). A salinity anomaly in the subpolar salinity distribution is applied to an idealized model of the Atlantic ocean to obtain the influence of a near-bottom transport on the sensitivity of the THC.

2 Model

The ocean model used is the GFDL primitive equation model MOM (Pacanowski et al., 1991). The domain of the ocean circulation model is a 70^0 wide sector from 66° S to 80° N including an idealized Greenland-Scotland ridge (compare Fig. 1). The model has a horizontal resolution of 2° and 15 vertical levels with grid size increasing with depth from 30 m to 836 m. The atmosphere model is a moist energy balance model (Lohmann et al., 1996 a; Lohmann and Gerdes, 1998). The atmospheric transport processes by transient eddies are modeled by diffusion.[1] Except for the near bottom transport parameterization, the coupled model is that of the sensitivity study of Lohmann and Gerdes (1998). In their study, the feedback mechanisms affecting the stability of the THC have been explored in the coupled atmosphere-ocean-sea ice system.

A slope convection scheme is introduced in the ocean model. In the parameterization, adjacent bottom tracer boxes are mixed at topographic slopes if dense water lies above light water on the slope

$$\nabla\rho \cdot \nabla h \quad > \quad 0 \qquad , \tag{1}$$

where ρ denote the density and h the topographic height, and ∇ the 3-D nabla operator. Technically, condition (1) is tested after the usual convective mixing algorithm and is applied successively to all pairs of layers, beginning from the top.

[1]The model is available under anonymous ftp at ftp.dkrz.de [136.172.119.129] /pub/Outgoing/gerrit/EBM

3 Experiments

Two set of experiments are performed, one without the new parameterization (experiment A) and one (experiment B) with the slope convection scheme.

3.1 Reference states

The overturning stream function of the reference state without slope convection (Fig. 1) is that from Lohmann and Gerdes (1998). The deep water formation takes place north of 50^0 N and the maximum overturning rate is 24 Sv. The mass exchange with the idealized Nordic Seas north of the ridge at $66°$ N is only 1 Sv. Here, we want to concentrate on the northern source of the THC and exclude any southern hemispheric deep water formation. The model is forced by fixed wind stress which was obtained from the Hellerman and Rosenstein (1983) climatology. The stream function of the vertically integrated transport for experiment (A) is shown in Fig. 2.

With the new parameterization (B), the meridional overturning in Fig. 3 is quite similar to experiment (A). The deep water is formed partially on the sill, north of the deep water formation areas of experiment (A). In (B), the deep western boundary current, the subpolar and subtropical gyres are enhanced (Fig. 4) compared to the barotropic mode of (A) shown in Fig. 2. The changed barotropic mode for the reference states in (A) and (B) can be understood by the barotropic vorticity balance. For both experiments, we have used the same wind stress, and the intensification of the horizontal gyre in (B) can therefore be attributed to the changed density structure in the joint effect of baroclinity and relief term (Sarkisyan and Ivanov, 1971; Mertz and Wright, 1992) and to frictional effects south of the sill.

Due to the changed circulation in the subpolar gyre, the surface water in (B) is fresher north of $40°$ N compared to (A). Along with the fresher surface water, the maximum overturning in (B) is slightly reduced for the equilibrium state (Fig. 3) compared to (A).

3.2 Perturbation experiments

Both systems are perturbed in surface salinity by -1.5 psu between $50°$ and $66°$ N in the top layer. This effectively shut down the source of deep water formation south of the sill. Fig. 5 shows the anomalous overturning stream function in the northern hemisphere for both experiments. The figure reveals a notable difference for the experiments (A) and (B). The system with the new parameterization recovers completely after 100 years, whereas (A) settles into a new equilibrium without North Atlantic Deep Water formation (case "3x" in Lohmann and Gerdes, 1998). On the long time scale, the model (B) shows a distinct reduction of overturning sensitivity to the negative salinity perturbation. We conclude that the new parameterization provides a stabilizing effect for the circulation with respect to subpolar sea surface salinity perturbations. Without effects of the new parameterization, we would have expected that system (A) is more stable than (B), because the THC is the more sensitive the weaker the thermohaline overturning is (Lohmann et al., 1996 b; Prange et al., 1997; Tziperman, 1997).

To understand the different behavior of (A) and (B), we analyze the anomalous sea surface salinity and temperature after the imposed perturbation averaged between 50^0 and 60^0 N. The time series (Fig. 6) show that the response of (B) is much smaller for sea surface temperature and salinity than (A). Furthermore, the reduction in surface heat loss (Fig. 6 c) due to the freshening is less than 40% of the reduction in (A). The stream function for the zonally integrated mass transport 10 years after the initial perturbation are shown in Fig. 7 for both experiments. This figure shows that deep water formation is interrupted in case (A) whereas in (B) deep water formation can continue. In the lower panel, we find that transport of deep bottom water takes place on the sill.

To analyze the dynamical effect of the slope convection scheme, temperature and velocity in the bottom layer are shown together (Figs. 8, 9). Fig. 8 shows the temperature and velocity in the bottom layer for both experiments before the perturbation has been induced. The water mass in (B) is coupled across the sill, and the deep water is colder south of the ridge and warmer north of the ridge compared to (A, Fig. 8). Moreover, the zonal transport in the bottom layer is enhanced in system (B). Fig. 9 shows the situation 10 years after the perturbation. Deep water production is interrupted in (A) and the bottom water south of 50° N has become warmer. Between 50° and 60° N, the upper panel in Fig. 7 shows an intermediate water production, corresponding to strong westward velocities in the bottom-most layer at 57° N. In experiment (B), the meridional overturning (lower panel in Fig. 7) is reduced, but can continue. This corresponds to relatively cold bottom water and strong westward velocities in the bottom layer (lower panel in Fig. 9).

4 Discussion

A large salinity perturbation at the high latitude surface weakens the THC (Fig. 5) for both experimental sets (A) and (B). In the experiment with the slope convection parameterization, the convectively produced water north of the sill can contribute to deep water formation and the dense signal of relatively cold water can reach the bottom (lower panel in Fig. 8). This cold water mass flows southward onto the ridge at the eastern boundary, and follows the ridge zonally at the southern slope (Figs. 8 and 9) spreading westward. This corresponds to the overflow pathway outlined by Gerdes (1993 b).

The added fresh water prevents vertical mixing which is the important process on the local scale affecting the large-scale meridional overturning (Lohmann and Gerdes, 1998). In (A), the freshwater cap prevents vertical mixing because the cold and fresh surface water is not dense enough to reinitiate deep convection. In this experiment, the initial salinity perturbation of -1.5 psu in the surface layer (Fig. 6 a) induces a shallow meridional overturning (Fig. 7, upper panel). The anomalous temperature (Fig. 6 b) does not compensate for the density reduction due to salinity (Fig. 6 a).

The water north of the sill is not affected by the initial surface freshening. In (B), this water contributes to deep water formation and makes vertical mixing still possible. The relatively fast, bottom-intensified outflow (Fig. 9) entrains ambient water as they plunge south of the sill. By this process, dense water is entailed across the ridge stabilizing the zonal density gradients and thus the THC (lower panel in Fig. 7). This overflow process feeds back to the surface heat loss which is reduced due to the freshening (Fig. 6 c). The reduction in surface heat loss is much smaller in (B) than in (A) because vertical mixing still exists with the slope convection parameterization.

The local difference in slope convection and surface heat flux contributes to the large-scale THC feedback which determines the difference in meridional overturning, 10 years after the perturbation (Fig. 7) and its long-term behavior (Fig. 5), for the two experiments. Our experiments suggest that the THC is less sensitive to surface salinity perturbations in the presence of a northern source of deep water and a near bottom tracer transport, than in the usual z-coordinate models with insufficient coupling across the sill.

Döscher and Redler (1997) analyzed the relative influence on meridional overturning of the formation of different North Atlantic deep water masses. They employ restoring of temperature and salinity to climatological values at the surface and at lateral sidewalls including the Denmark Straits at their northern model boundary. Their experiments suggest that the THC strength is much more sensitive to forcing at the northern sidewall than to surface forcing. A switched off surface forcing reduces the overturning by only 1 Sv and plays therefore a minor role driving the overturning. With no restoring in the northern boundary zone, the overturning reduces from 17 to 11 Sv and the surface forcing becomes important. Döscher and Redler (1997) therefore suggested that models without an adequate overflow could possibly be unrealistically sensitive to variable surface fluxes in the North Atlantic.

In our model, it is shown for the first time that the sensitivity to subpolar surface forcing is strongly reduced, if dense water at the northern sill can descend in a more appropriate way. Moreover, a better coupling between the North Atlantic and the Nordic Seas is accomplished in the model. These improvements can be achieved by using a simple slope convection scheme which distinctly reduces the artificial dilution of the water mass at the continental slope.

Our model in the simplified configuration used here has the advantage of simplicity because of neglected topographic details and the atmospheric response model. It is believed that the moist energy balance model captures some basic features which are necessary for the interactions with the oceanic thermohaline circulation (Lohmann and Gerdes, 1998). As in other studies (Gerdes, 1993 b; Beckmann and Döscher, 1997; Winton et al., 1997; Gnanadesikan et al., 1998; Killworth and Edwards, 1998), an idealized ridge in the model is useful to obtain the mechanism of the new parameterization with respect to the sensitivity. In the experiments shown, we find that the overflow is too weak, which is a problem of the idealized model geometry chosen and of insufficient water mass transport from the north. The logical next step would be a model study with more realistic bottom topography, and including water masses from the Mediterranean and the southern hemisphere.

5 Conclusions

Analyzing the sensitivity of the THC with respect to high latitude surface forcing, the concept of a bottom boundary layer model is introduced in a coupled atmosphere-ocean-sea ice model. In the ocean model, the down-slope flow of dense water is parameterized as a slope convection scheme in a z-coordinate model, as proposed by Beckmann and Döscher (1997). Our ocean model with an idealized ridge in the northern North Atlantic does not feel the step like topography in the z-coordinate model. In this type of model, the dilution of water mass properties along the slope is reduced and cold surface water can reach the bottom. Although the slope convection is only a simple parameterization which does not involve additional boundary layer physics such as turbulence near the floor (Mellor and Yamada, 1974), it is capable to drastically improve the notorious disregard of physics of descending dense water plumes in coarse resolution z-coordinate-models. It is therefore believed that the mechanisms described in this note should be basically found when applying other and more sophisticated bottom boundary layer schemes (Gnanadesikan et al., 1998; Killworth and Edwards, 1998), because such schemes can adequately model descending dense water plumes.

A large negative sea surface salinity perturbation south of the ridge reduces deep water formation in the northern North Atlantic and leads to a decrease of high latitude temperatures. Such a sensitivity study is motivated by paleoclimatic cooling events due to fresh water events in the Labrador Sea (e.g. Fairbanks 1989; Dansgaard et al., 1993). With the slope convection parameterization, the system is not as sensitive to the subpolar freshening as in the z-coordinate system because slope convection allows a coupling over the sill. Deep convection at subpolar latitudes is no longer the dominant forcing for the THC if a northern source of deep water formation is present. Freshening events in the Labrador Sea may therefore affect deep water formation not in the way suggested by models without transport of dense water on the slope.

For the present climate, there are no indications for changes in the overflow water over the Greenland-Scotland-Ridge (Saunders, 1990; Dickson and Brown, 1994). It is likely that this is a strong stabilizing mechanism for the deep water formation under Holocene conditions. It is a question why the present day's overflow remains constant (Saunders, 1990; Dickson and Brown, 1994) despite interruptions of deep water in the Nordic Seas (Schlosser et al., 1991). Contrary to the Holocene conditions, the THC's sensitivity against surface forcing is probably increased in the glacial climate because of a reduced overflow. A narrower opening at glacial times would restrict both the inflow of warm Atlantic water and the outflow of low salinity water from the Nordic Sea. Experiments of Redler and Böning (1997) using a high resolution ocean model south of the Greenland-Scotland ridge show that the deep western boundary current is sensitive with respect to the inflow points and the representation of the topographic ridge. It is conceivable that the sensitivity against surface forcing is increased in the glacial climate because of narrower openings and a reduced overflow. The increased sea ice cover in the northern North Atlantic at glacial times is a further destabilizing effect for the THC because ice suppresses vertical mixing effectively (Lohmann and Gerdes, 1998).

Our experiments suggest that without a sufficient coupling across the sill, the ocean's response to subpolar atmospheric variability may be overestimated. The reliability of climate experiments, which examine the response to high latitude forcing, would be improved if the different contributions to the deep water circulation south and north of the North Atlantic sill are taken into account. The representation of the downslope transport of dense overflow water seems to be important for estimating paleoclimatic and decadal climate variability.

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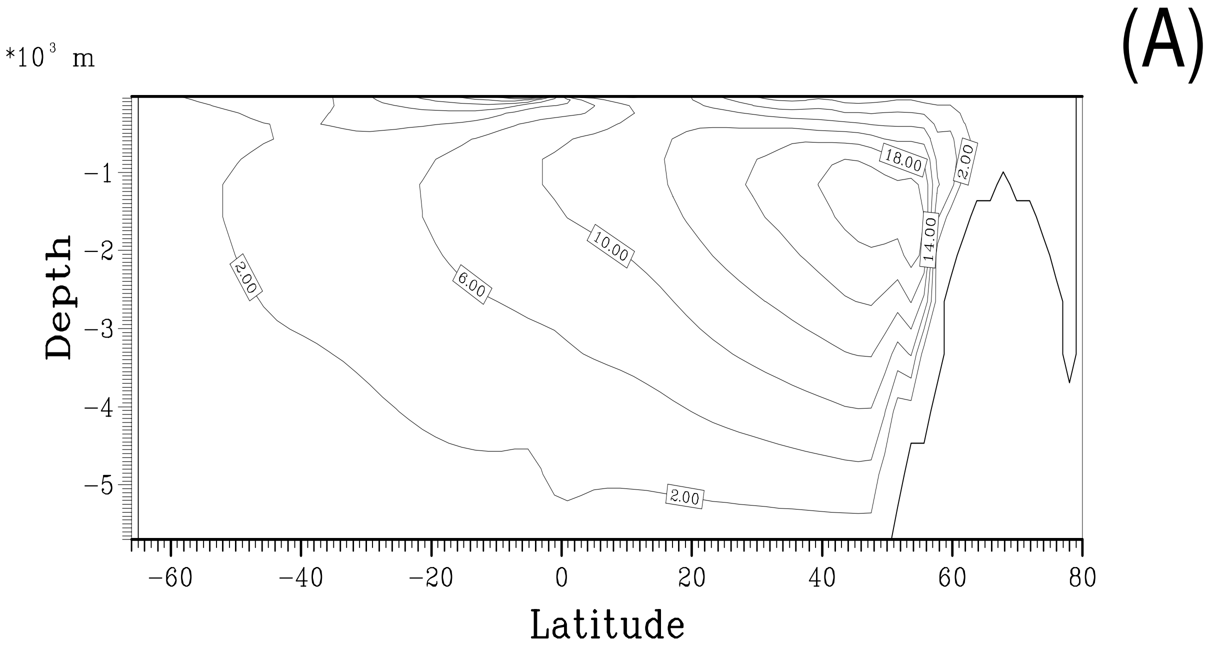

Figure 1: Reference stream function for the zonally integrated mass transport for experiment (A). The maximum overturning rate is about 24 Sv ($1Sv = 10^6\, m^3 s^{-1}$).

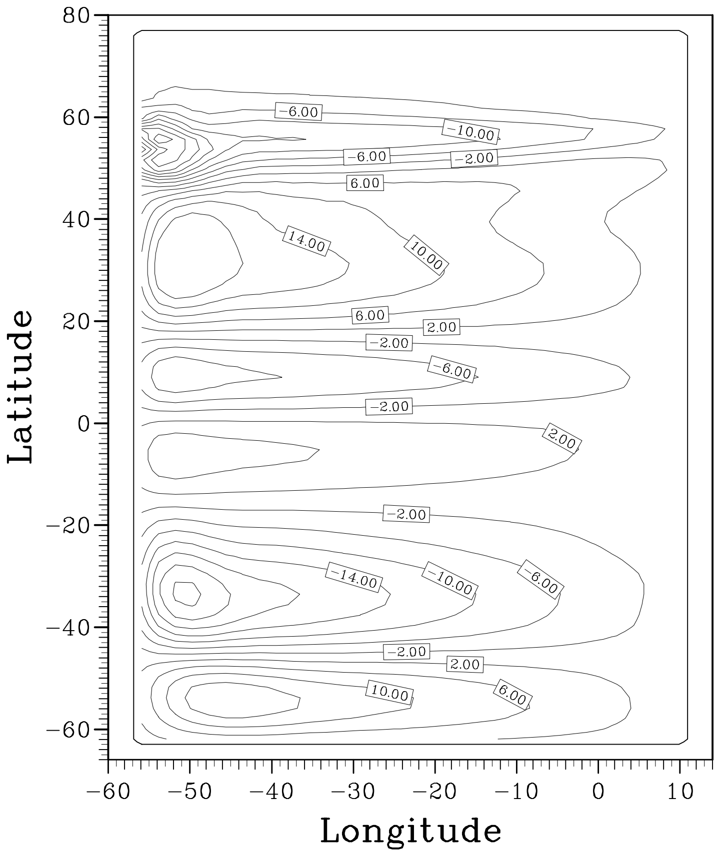

Figure 2: Stream function of the vertically integrated transport for experiment (A). The contour interval is 4 Sv.

(B)

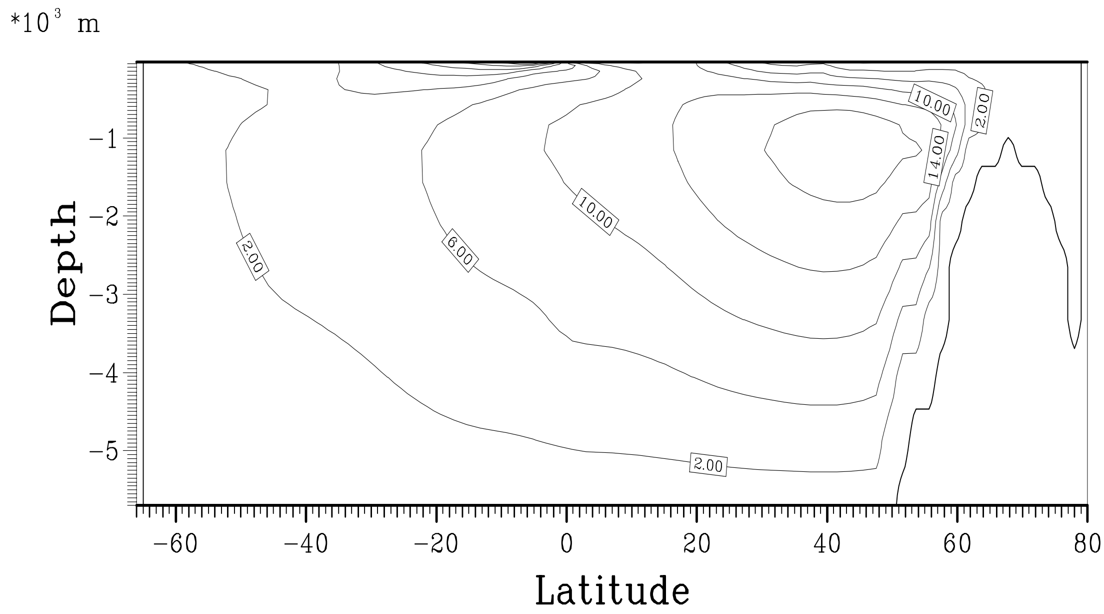

Figure 3: Reference stream function for the zonally integrated mass transport for experiment (B) with the new parameterization. The maximum overturning rate is about 20 Sv.

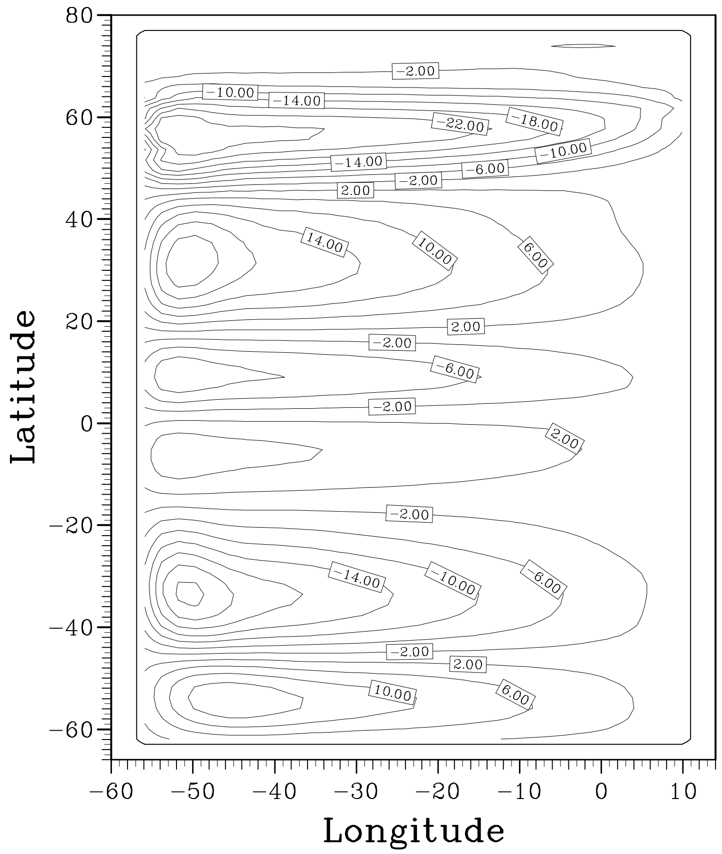

Figure 4: Barotropic stream function for experiment (B).

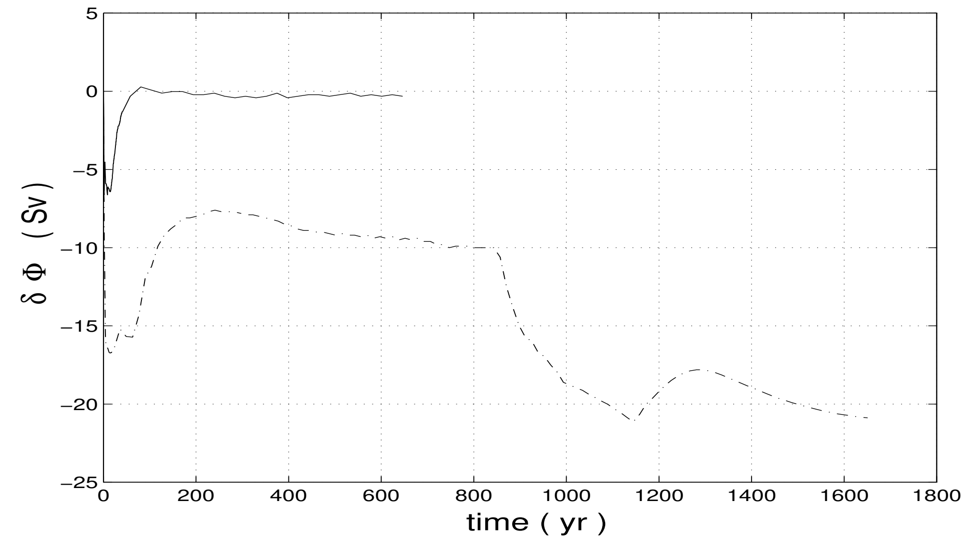

Figure 5: Change of the maximum overturning rate in the northern hemisphere for the experiment (A) (dashed dotted line) and experiment (B) using the slope convection parameterization (solid line). The unit is $Sv = 10^6\, m^3\, s^{-1}$.

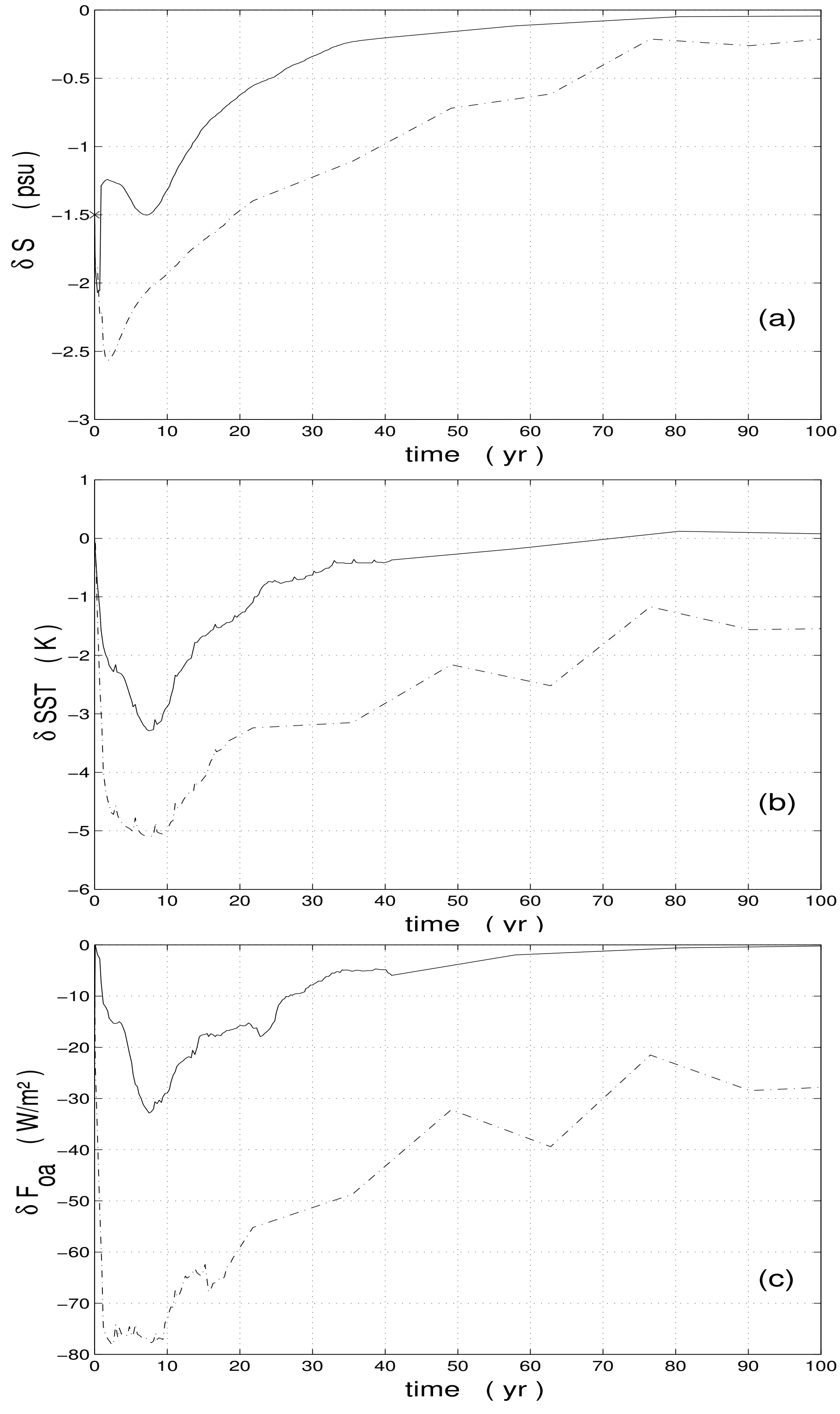

Figure 6: Time series of the anomalous sea surface salinity (a), temperature (b) and heat flux at the atmosphere-ocean interface (c). The values are averaged between 50^0 and 60^0 N. Experiment (A) has dashed dotted lines, experiment (B) with the new parameterization has solid lines. Both systems are perturbed at time zero in high latitude salinity by -1.5 psu (compare panel a).

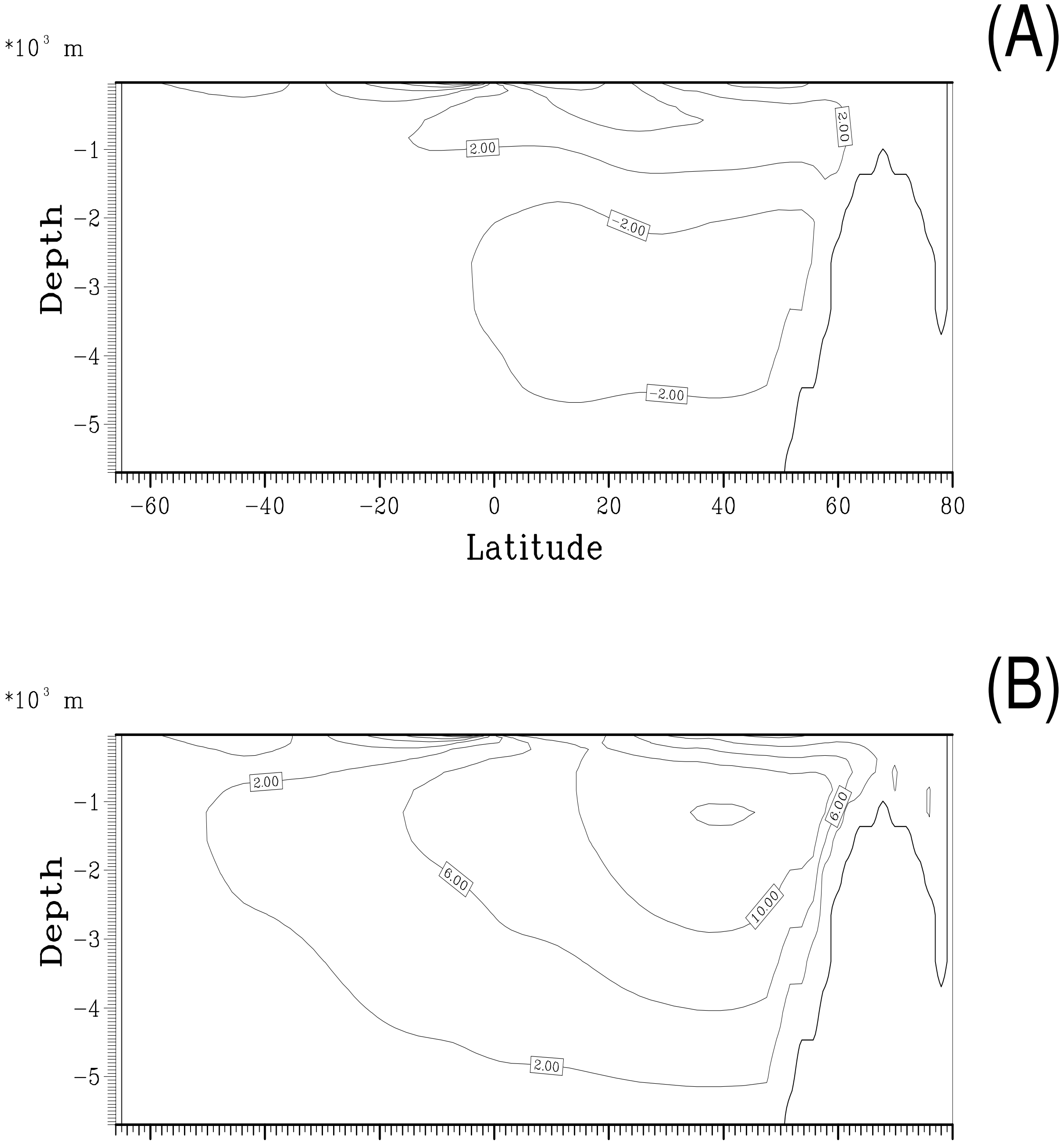

Figure 7: Stream function for the zonally integrated mass transport for (A) and (B), 10 years after perturbation. The contour interval is 4 Sv.

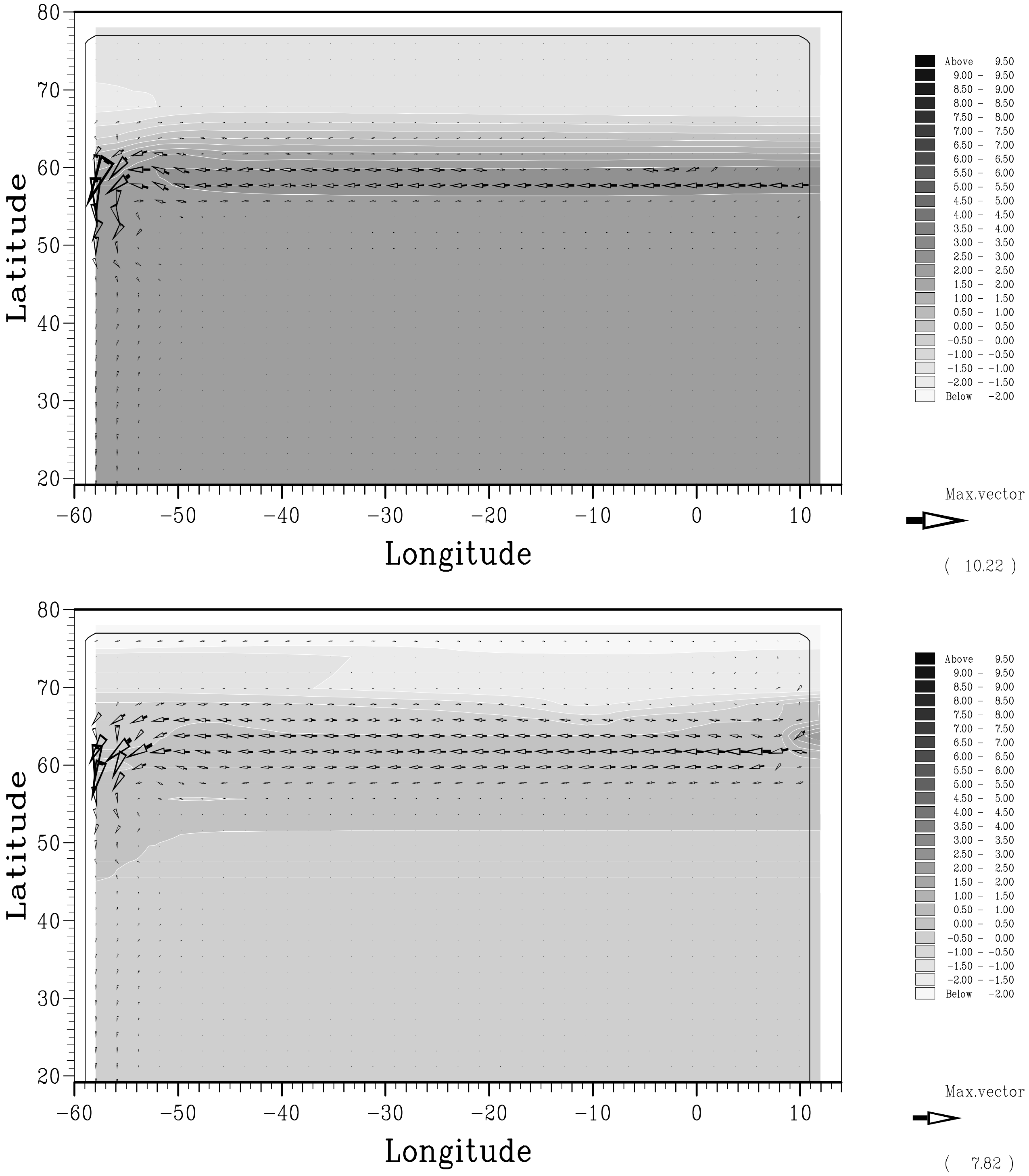

Figure 8: Temperature and velocity in the bottom layer for experiment (A, upper panel) and (B, lower panel) before the salinity perturbation has been induced. Units are $^{\circ}C$ and $cm\, s^{-1}$ with the same scaling for both panels.

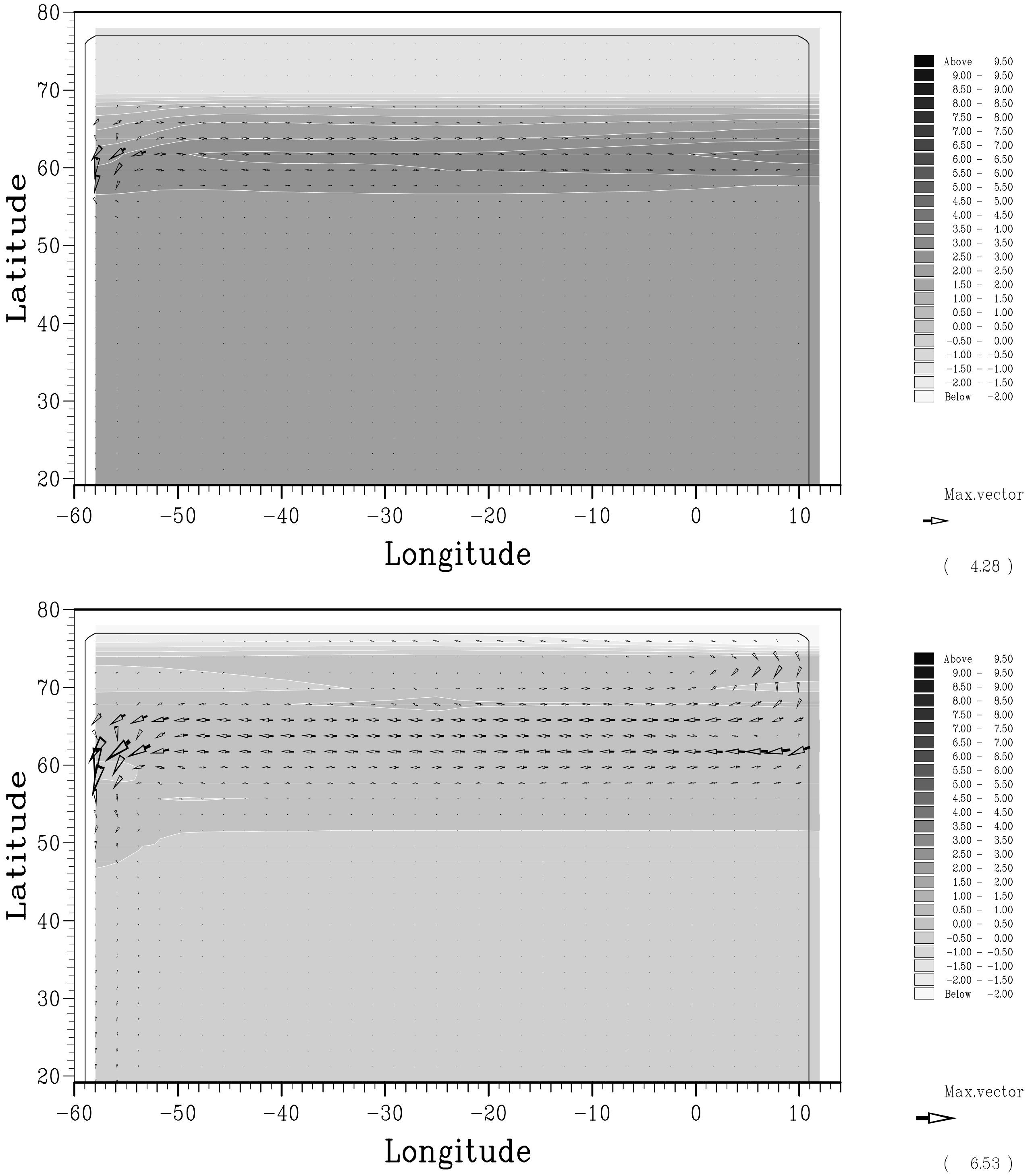

Figure 9: As in figure 8, but 10 years after the perturbation in sea surface salinity.