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Variability on the century time scale and regime changes in a stochastically forced zonally averaged ocean-atmosphere model

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Abstract. The response of a zonally averaged climate model of the atmosphere-ocean system to a stochastic freshwater forcing depends on its strength. Three regimes are found: For small perturbations of the system, the ocean responds with irregular oscillations around a state which is characterized by a strong Atlantic overturning circulation. For intermediate to large amplitude forcing the thermohaline circulation in the Atlantic ocean exhibits pronounced regular overturning (flushing) events. The third regime is a complete collapse of the Atlantic overturning. Transitions can be induced between the three regimes. While for moderate values the transitions are reversible, for large forcing amplitudes the Atlantic overturning circulation collapses irreversibly to the third regime.

Introduction

Natural variability of the thermohaline circulation on decadal and longer time scales is still rather poorly understood. The Atlantic's thermohaline circulation is responsible for the major part of the oceanic heat transport from low to high latitudes. The flow is driven by surface heat and buoyancy fluxes. Sea surface temperature (SST) and sea surface salinity (SSS) variations, especially in high latitudes, have a direct influence on the role of deep water formation and thereby on the entire overturning circulation.The surface freshwater balance is an important control variable and thus, a stochastic model is a good first approximation to study its effect on the ocean [*Hasselmann*, 1976]. This idea has been realized in a number of studies that employed three-dimensional [*Mikolajewicz and Maier-Reimer*, 1990; *Weisse et al.*, 1994; *Weaver et al.*, 1993; *Capotondi and Holland*, 1996], two-dimensional ocean circulation models [*Mysak et al.*, 1993] and also box-models [*Bryan and Hansen*, 1995; *Griffies and Tziperman*, 1995].

The purpose of this study is to gain some understanding of the mechanisms involved in the long-term variability of the ocean's thermohaline circulation and to explore their role in triggering transitions between different steady states. The present study is a continuation of the work by *Mysak et al.* [1993] who performed stochastic forcing experiments using an ocean-only, one-basin geometry version of the model of *Wright and Stocker* [1992].

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The model set up

We use the two-dimensional, zonally averaged 3-basin ocean circulation model of *Wright and Stocker* [1992] with a new closure scheme [*Wright et al.*, 1998], coupled to a one-dimensional, zonally and vertically averaged energy balance model of the atmosphere [*Stocker et al.*, 1992] with an active hydrological cycle [*Schmittner and Stocker*, 1999]. Detailed model description and geometry is given by *Schmittner and Stocker* [1999].

The model is first spun up from rest for 7,000 years by relaxing temperature T and salinity S in the uppermost layer on a time scale of 50 days to zonally averaged values of temperature and salinity [*Levitus et al.*, 1994; *Levitus and Boyer*, 1994], zonal wind stress is taken from *Han and Lee* [1983]. At year 7,000 the vertical salt flux at the surface is diagnosed and then used as the boundary condition for S during a further 3 kyr integration period. At year 10,000 the atmospheric parameters are determined and the model is integrated in the coupled mode for another 7,000 years to reach steady state.

For the next 18,000 years a random freshwater flux is superimposed on the basic-state freshwater flux. The random component consists of a spatially uncorrelated, zero-mean Gaussian white noise process which has standard deviations ranging from 0.025 Sv to 0.13 Sv (in a single run all ocean surface boxes are subject to the same, total freshwater amplitude).

The model response

Except for very large forcing, where the Atlantic overturning cell collapses, the model displays two different types of oscillations. These have characteristics which can be found in all observed quantities, such as meridional heat flux or Atlantic overturning. In both cases (in the following termed type A and type B oscillations) fluctuations occur dominantly on the century time scale and appear strongest in the Atlantic Ocean. Oscillations in the Pacific and Indian Ocean are highly correlated with those in the Atlantic, indicating that the Southern Ocean strongly couples the ocean basins.

In type A oscillations there are additional variations on a time scale of a few decades, while in type B we only observe variations on a century time scale (Fig. 1, top). Type B oscillations are similar to the flushing cycles described by *Weaver and Hughes* [1994], although here they evolve on a shorter time scale.

The power spectrum of the heat flux across 32.5°N (Fig. 1, bottom) exhibits a peak at 176 years and additional power in the band of 20 to 50 years. For type B oscilla-

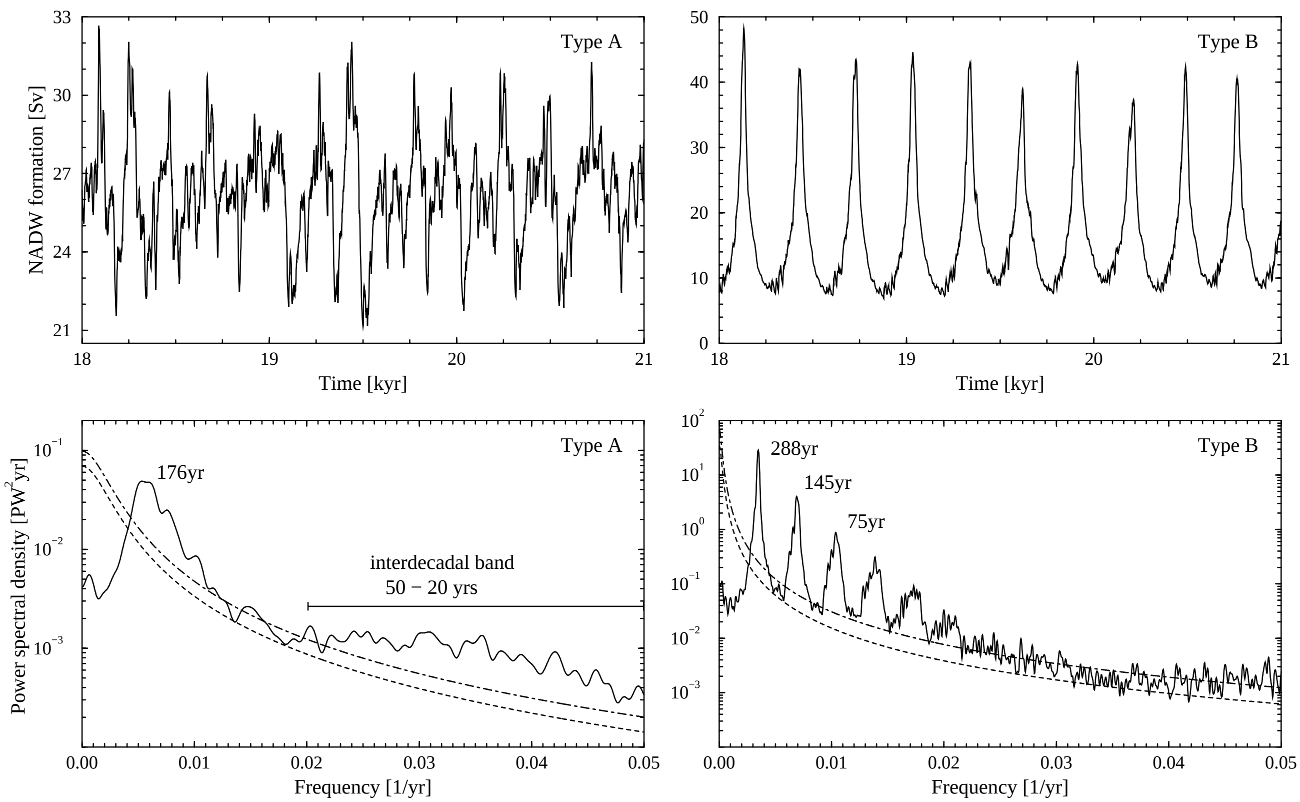

Figure 1. Atlantic overturning (top) and their spectral estimates (bottom) for type A and type B oscillations. The solid curve is a Blackman-Tukey estimate of the power spectrum with a Bartlett window. The dashed curve is a red-noise model and the dash-dotted curve shows the 95% significance band.

tion the peak at around 288 years can be identified with the clearly visible period in the time series of the NADW formation (Fig. 1, top right). Type A has a bigger mean overturning than type B. This is due to a relatively strong negative density anomaly which develops north around 40°N for type B. Overturning is also shallower in type B and the deep ocean is significantly colder due to a dominance of waters of southern origin below 2500 m depth.

The nature of the oscillations can be further analyzed by performing a principal components analysis of the stream function (not shown). For both oscillation types, the first eigenvector represents a basin-scale variation of the mean circulation pattern. The Southern Ocean as well as the equatorial Ekman cells do not participate in the oscillation. This pattern accounts for about 85% and 95% of the total variance for type A and type B, respectively.

Salinity, temperature and density fields were analyzed using the Principal Oscillation Pattern (POP) method [*Von Storch and Navarra*, 1995]. Both, A and B modes have a similar oscillatory mode in salinity anomalies with periods of 200 and 290 years, respectively. A negative salinity anomaly is developed at around 40°N which is balanced by an extended anomaly north and south of the equator (Fig. 2) in the Ekman cells of the mixed layer. The thermohaline circulation advects the positive tropical salt anomaly to the sinking region at 60°N which then increases the convection and thereby the thermohaline circulation. The positive anomaly is "flushed" downward, and a strong basin-scale overturning is developed. As soon as the water column is stabilized (by partial or total elimination of the fresh water pool) the overturning reduces, and stronger wind-driven cells develop which again start to capture the freshwater input of the atmosphere.

Oscillation regime changes

Transitions between different oscillatory states can be induced by changes in the strength of the random forcing. In a first experiment, starting from the steady state a stochastic component is added to the freshwater flux forcing with a strength equivalent to 0.028 Sv. During the following 70 kyr the strength is increased by 10^{-3} Sv every 2,000 yr (Fig. 3 top); from year $87,000$ onwards the strength is decreased stepwise. For the first 57 kyr, the model shows type A oscillations, but when the forcing strength exceeds 0.055 Sv the model shifts rapidly from type A to type B oscillations. These oscillations are maintained until the strength of the forcing is reduced below 0.045 Sv. At this point, the model shifts back to type A oscillations. This state persists during the rest of the integration. These regime changes are reversible and the fact that the transition point depends on the oscillation state prior to the transition, implies hysteresis behavior as a function of the forcing strength.

In a second experiment (Fig. 3 bottom), we increased the forcing from 0.025 Sv to 0.13 Sv. The model again shows a transition from type A to type B oscillations when the forcing exceeds 0.055 Sv. The system is approaching the stability limit which is evident from the slow growth of the

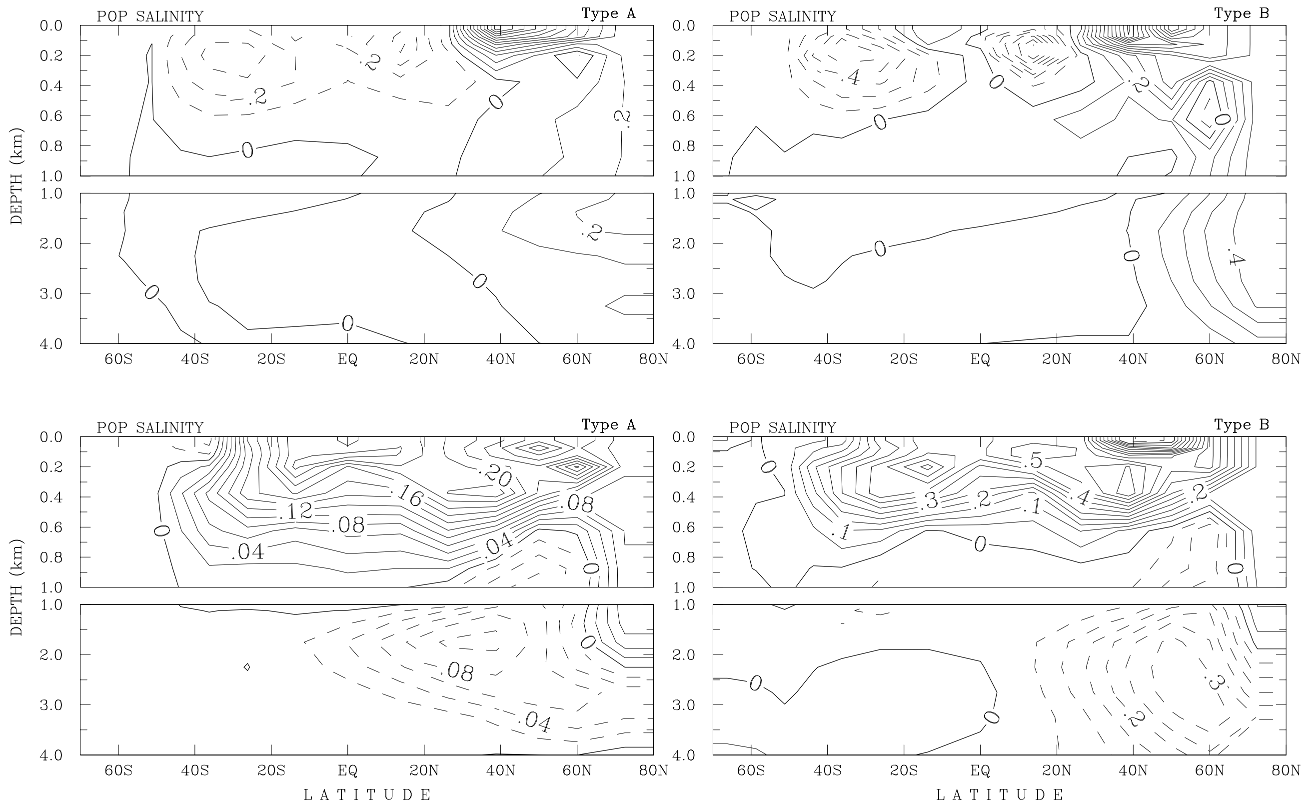

Figure 2. First Principal Oscillation Pattern of the salinity in the Atlantic for type A (left column) and type B (right column) oscillations. The top/bottom row displays the real/imaginary part.

amplitude of the maximum overturning in the Atlantic. At 0.115 Sv NADW formation shuts down within about one hundred years. In turn, the Pacific establishes deep water formation in the high northern latitudes. This new circulation pattern is maintained even when the forcing decreases to its initial value or is removed completely, i.e. here the collapse of the Atlantic overturning is an irreversible process. Instead, the Pacific now forms a positive overturning cell with deep water formation in the high northern latitudes, and a reversed conveyor belt establishes.

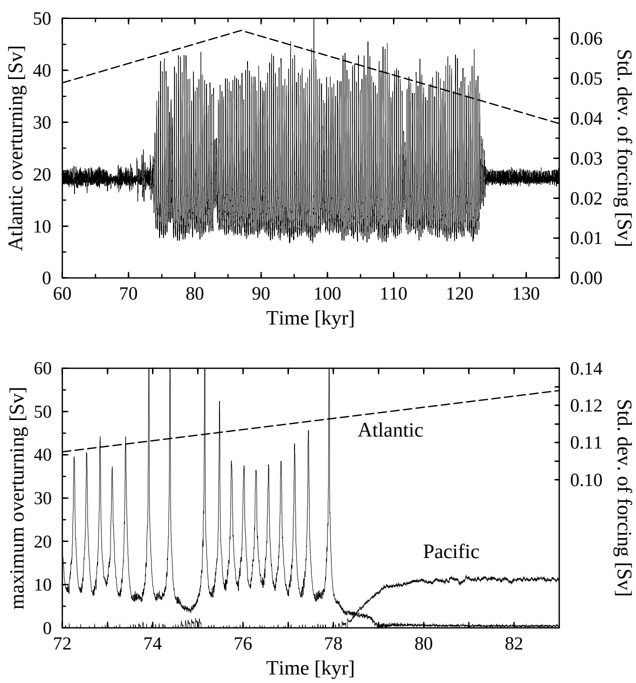

Figure 3. Top: Regime changes found in time series of the Atlantic overturning and the forcing (dashed line).

The distinction between the individual regimes can be clearly shown by performing long runs for a wide range of forcing amplitudes and measuring the mean and the standard deviation of the maximum Atlantic deep overturning (Fig. 4). Although the mean circulation for both oscillatory states is similar, its variability as indicated by the standard

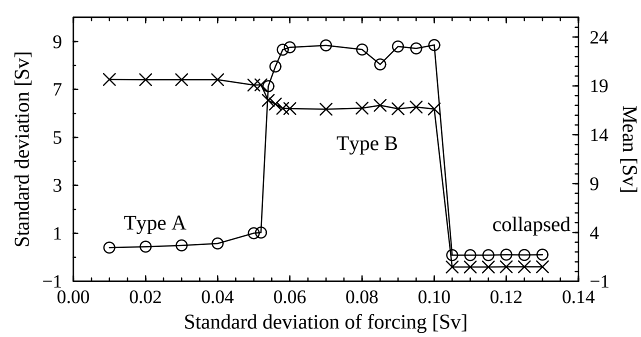

Figure 4. Mean (dashed line) and standard deviation (solid line) of the maximum deep overturning in the Atlantic as a function of the standard deviation of forcing. Each point represents an individual experiment with a fixed strength of the random component of the freshwater forcing.

deviation, changes almost by a factor of 10. For amplitudes larger than the equivalent of 0.12 Sv the Atlantic ocean has a very small negative overturning cell with little variability.

Increasing random forcing thus causes the crossing of a threshold value below which the overturning circulation can no longer be maintained. The new equilibrium state is significantly more stable; no strong fluctuations are observed even though freshwater perturbations continue to increase.

Conclusion

The one-basin experiments of *Mysak et al.* [1993] showed decade to century time scale oscillations. Using a more realistic geometry (3 basins), a better parameterization of the dynamics [*Wright et al.*, 1998] and an atmospheric energy balance, we found variability on similar time scales of 200 - 300 years. It has been argued that the thermohaline circulation may be responsible for millennial fluctuations found in paleoclimatic records [*Bond et al.*, 1997]. The present model results suggest that the range of 1,000-2,000 years is beyond the natural time scales of the Atlantic thermohaline circulation. However, changes in the strength of the hydrological cycle caused, for example, by slow changes in the Earth's orbital parameters or low-latitude natural climate variability, may well trigger regime changes of the thermohaline circulation.

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