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VARIABILITY OF DEEP WATER FORMATION AND CONVCTION IN THE NORTH ATLANTIC: A MODEL STUDY

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1. INTRODUCTION

The lower branch of the thermohaline circulation in the Atlantic Ocean is driven by two sources: the convection in the Labrador Sea (LS) and the overflows across the Greenland Scotland Ridge (GSR), which are fed by deep water formation processes in the Greenland Iceland Norwegian (GIN) Sea and the Arctic Ocean. The overturning circulation therefore responds to variations in large-scale atmospheric features. Examples are the observed correlations between the North Atlantic Oscillation (NAO) and the properties of the Labrador Sea Water (LSW) (Curry et al., 1998). According to Dickson et al. (1996) the NAO also triggers the relative importance of the deep water formation regions to the north (GIN Sea) and to the south (LS) of the Greenland Scotland Ridge.

2. THE MODEL AND THE EXPERIMENTAL SET-UP

In the experiment presented here we study the ocean response to atmospheric variability using the C-HOPE OGCM (Maier-Reimer et al. 1997) newly developed at the MPI in Hamburg. The model includes the option of conformal grid mapping. The version applied here has a formal zonal resolution of 3˚, with the poles in central Greenland and Antarctica. The grid cells are quadratic with resolution ranging from 20km near the east Greenland coast to 350km in the tropical Pacific. The model has a free surface and includes a state of the art sea ice model with viscous-plastic rheology. The daily forcing in the simulation has been derived from the NCEP reanalysis data (Kalnay et al. 1996) of near surface air and dew point temperature, 10 m windspeed, downward shortwave radiation, precipitation, cloudiness and wind stresses. The heat fluxes are calculated from the atmospheric forcing data and the actual model distribution of SST and sea ice using bulk formula. For freshwater a mass flux boundary condition is implemented, where the actual flux is calculated from prescribed precipitation and climatological river runoff and evaporation calculated from the latent heat fluxes. Additionally a weak restoring of the surface salinity towards the Levitus (1994) climatology is used with a time constant of 386 days. The model has 20 levels, thicknesses increasing with depth from 20m at the surface to 1400 m in the bottom layer.
The model is spun up for 500 years using climatological forcing after starting from Levitus. Thereafter, the forcing is switched to NCEP data for the period 1948 to 2000 and applied two times subsequently. In the following we present results from the second run with the NCEP forcing.

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3. RESULTS

3.1 *Variability in Labrador Sea convection*

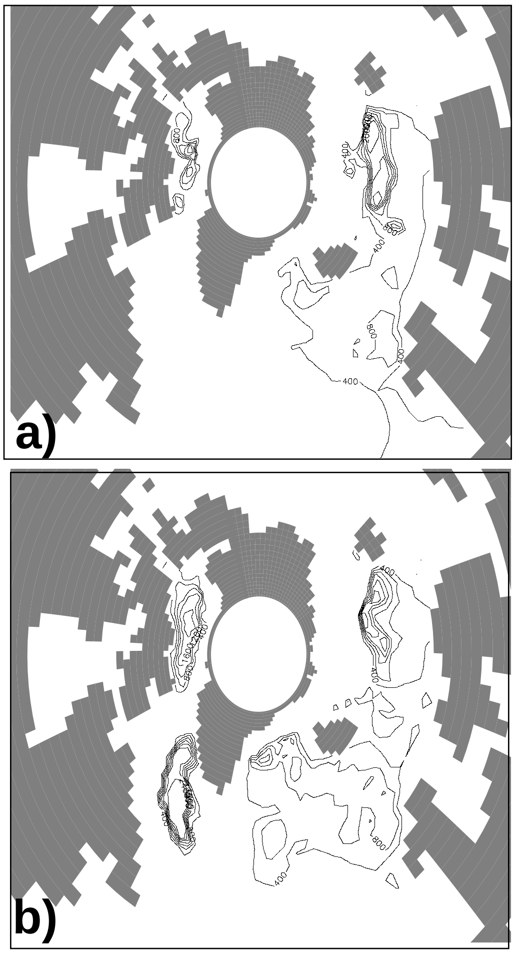

Figure 1: Depth of the winter mixed layer in a) March 1963, and b) March 1988

The model clearly simulates interannual and decadal variability. The relatively high resolution in the Labrador Sea and the GIN Sea enables us to simulate the convective activity in these two regions more realistically than previous coarse resolution models. The centers of convection are situated in the Greenland gyre at about 75^{o}N and in the central LS. Heat loss to the atmosphere is strongly variable and depends on the local surface conditions. In particular, we find a sensitivity of the convective activity to decadal atmospheric variability and, hence, to the NAO as the dominant mode in the North Atlantic area. During NAO low phase convection is reduced in the LS and enhanced in the GIN Sea, vice versa during NAO-high periods. As an example we show in Figure 1the mixed layer depths analyzed from monthly means in the late winter (March) of 1963 (Figure 1a) and 1988 (Figure 1b)

Almost no convection takes place in the LS in the mid-60s where the NAO index was extremely low whereas convective water masses penetrate down to more than 2000 m in the mid-90s. LSW forms the upper part of the North Atlantic Deep Water (NADW) and cools and thickens as a result of increasing convection. The intensity of convection in the LS is determined by the strength of the westerly winds and the occasional passage of freshwater surface anomalies (Curry et al., 1998).

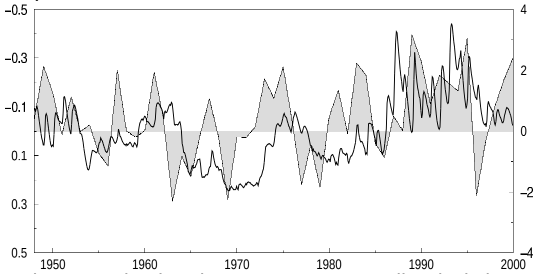

Figure 2: Simulated temperature anomalies (axis inverted, monthly means horizontally averaged over 9 gridpoints in the central LS) together with a time series of the NAO

Figure 2 shows a model time-series of potential temperature anomalies at 1400 m from the central Labrador Sea. There is a general cooling trend from the

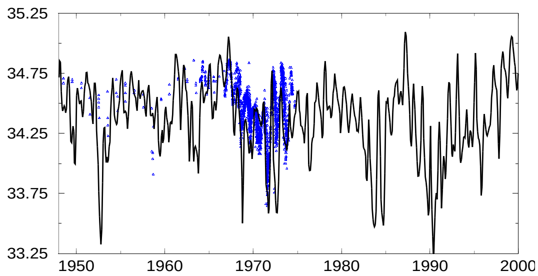

Figure 3: Near surface (30m) salinity (monthly average) from the simulation (solid line) together with observations from OWS Bravo (triangles). The model data were averaged over 9 gridpoints in the central LS.

mid-50s and 60s to the 80s and 90s, corresponding to the shift from the NAO-low regime to the NAO-high period in the last two decades of the 20th century. Our time series agrees well with Curry et al.'s (1998) observations from the central LS (not shown). In particular, the sequence of a temperature maximum around 1970, a minimum at 1975, and second maximum in the early 80s are reproduced. Curry et al. (1998) explain these positive events by the buildups of extremely low surface salinities, which inhibit overturning convection.

In the model we find similar fresh water anomalies and in Figure 3 we compare the salinity in the 2nd model level (30m) with the corresponding time series of the ocean weather ship (OWS) Bravo. The model is capable to simulate the freshening in the late 60s and early 70s. It successfully reproduces the extreme freshening in 1971. Thereafter the salinity values slowly return to normal. Unfortunately, the data available to us covered only the period up to 1974. In the model we see additional pronounced freshwater anomalies in the 80s and around 1990. These agree favorably with the observations reported in the review article of Belkin et al. (1998) (see also the references therein).

3.1.1 *Upstream sources of fresh water anomalies*

The event around 1970 coincides with the "Great Salinity Anomaly" (GSA) as described by Dickson et al. (1988). This and the later anomalies (Belkin et al., 1998) are advectively transported from the Arctic ocean via the East Greenland Current. In our simulation we find that they are related to extremely high amounts of Arctic sea ice transported to the south through Fram Strait. The sea ice volume in the Arctic shows a pronounced response to the imposed atmospheric variability. There are distinct maxima in the winters of 1965/66 and 1987/88 and smaller peak in the winter 1979/80 (not shown). In these years sea ice piles up in the Pacific sector of the Arctic. In the following years the ice volume shows a dramatic decrease of up to 5000 km^3. The sea ice is transported across the Arctic, leading to positive ice thickness anomalies close to northern Greenland.

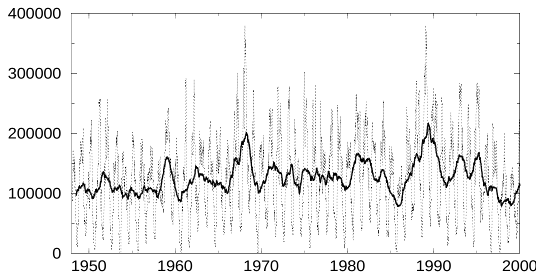

Figure 4: Ice transport (in m^3/s) through Fram Strait. Monthly means (gray line) and annual means (black line)

The time series of the sea ice export from the Arctic ocean through Fram Strait (Figure 4) shows that the decrease in Arctic ice volume is mainly caused by enhanced sea ice export: An enhanced export of 80,000 m^3/s averaged over 2 years is required to explain the observed reduction in Arctic ice volume. This enhanced freshwater supply to the Nordic Seas is transported southward with the East Greenland Current till it reaches the Labrador Sea. In the Labrador Sea a negative surface salinity anomaly develops and influences the convective activity there. The salinity signal of the early 70s GSA further propagates across the North Atlantic and reenters the Norwegian Sea via the Faroe Shetland Channel around 1975.

3.2 *Variability in The Denmark Strait Overflow*

The deeper layers of the NADW are fed by the overflows across the GSR through two principle pathways, the Faroe Bank Channel and the Denmark Strait. The Denmark Strait overflow produces the coldest and densest components (DSOW) of the NADW. The flow through straits is constraint by hydraulic conditions and depends on the upstream conditions (e.g. Käse and Oschlies, 2000). The latter are subject to changes in the water mass distribution and variations in the deep and intermediate water formation in the Nordic Sea.

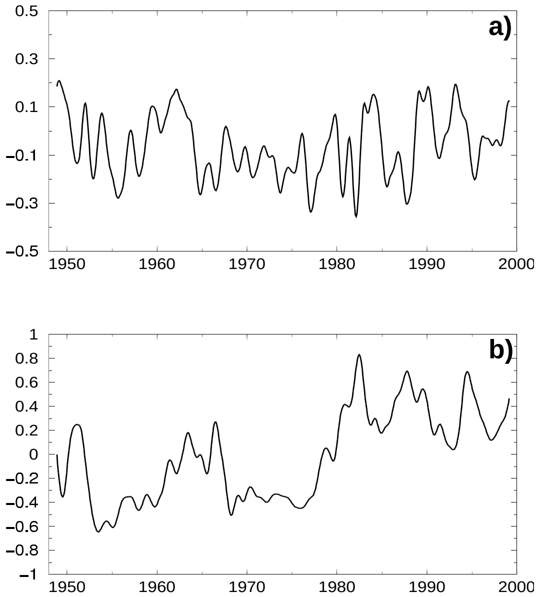

Figure 5: Variability in the Denmark Strait overflow. Anomalies of a) transport (units: Sv, 1 Sv = $10^6 m^3/s$) southward transports below 400m with temperatures below 2.5C), and b) temperatures (averaged over the outflow layer below 400m)

Figure 5 depicts a timeseries of transport anomalies (Figure 5a) and temperature anomalies (Figure 5b). Whereas the transport fluctuations are relatively small (less than 10% of the mean) there are pronounced variations in the temperature. The overflow is generally warming throughout the 80s, thus showing an opposite trend compared with the deep LS (Figure 2). This reflects the general trend to less active convection in the Nordic Seas. However, the DSOW is not directly linked to the deep convection in the Greenland Sea gyre. Rather, it is a blend of water masses that are modified in the Norwegian Sea, the Arctic Ocean and the Greenland and Icelandic Seas (Rudels et al., 1999). There is also evidence that the recipe of the DSOW mixture is changing with time. In our analysis we found, for example, that the relatively low overflow temperatures in the early 70s are related to unusually intense intermediate-depth convection in the Iceland Sea. At this time extremely cold conditions, leading to increased ocean-atmosphere heat fluxes, were observed in the Jan Mayen region (Bacon, 1998).

These fluctuations are then quickly and relatively undiluted transferred to Denmark Strait where they are imprinted onto the DSOW and subsequently onto the NADW.

Our ongoing analysis shall focus on the downstream fate of the polar and sub-polar climate signals in the North Atlantic.

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