The simulation presented here takes place in the framework of the PREDICATE project. This research program's central objectives are (a) to improve understanding of the processes responsible for decadal-timescale fluctuations in the climate of the Atlantic-European region, and (b) to assess the extent to which such fluctuations are predictable. The coupled model used to answer these questions is the ARPEGE3/ORCALIM2 model. A 200-year simulation has been performed without flux correction or restoring terms, which will help to understand the mechanisms underlying multi-decadal fluctuations in the climate system. Two 25-year long ensembles of predictability experiments are conducted from this simulation, contrasting two phases of the low frequency fluctuations of the thermohaline circulation (THC). These forecasts will help us in determining the potential predictability of decadal fluctuations in the North Atlantic Europe region. The first part of this report describes the models and experimental set-up. The second part gives some insight on the model's variability in the North Atlantic Europe region.

1 Model description and experimental design

1.1 The atmospheric component

The Atmospheric Global Circulation Model (AGCM) used in this atlas is the third version (referred as to cycle 18) of the ARPEGE-Climat model developed at Météo-France from the ARPEGE/IFS operational weather prediction model jointly maintained by the European Center for Medium range Weather Forecasts (ECMWF) and Météo-France. The ARPEGE model is a spectral model and uses a two time level semi-lagrangian scheme with semi-implicit time discretization of the equations. The standard configuration of the climate version employs a T63 triangular horizontal truncation. Diabatic fluxes and non-linear terms are calculated on a gaussian grid of about 2.8° degrees longitude by 2.8° degrees latitude. The vertical is discretized over 31 levels (20 levels in the troposphere) using a progressive vertical hybrid coordinate extending from the ground up to about 34 km (7.35 hPa). The model time step for this resolution and time discretization scheme is 30 minutes.

ARPEGE-Climat includes all basic atmospheric physical parameterizations as well as comprehensive land surface processes via the ISBA model (Douville, 1998). The convection is represented by a mass flux scheme with detrainment as proposed by Bougeault (1995). The radiative package follows the Fouquart and Morcrette's scheme (Morcrette, 1990). The longwave radiation parameterization includes the effect of trace gases (CH_4, N_2O, $CFC_{11\text{-}12}$) as well as CO_2, O_3, H_2O. Parameterizations of boundary layer and turbulence are based on vertical exchange coefficients computed as functions of the local Richardson number according to the Louis et al. (1982) method.

The major differences between the latest version of ARPEGE-Climat and the previous one (Déqué et al., 1994) include changes to the time discretization scheme as well as to the parameterization of cloud distribution and properties, deep convection and land processes. The stratiform and shallow convection cloud formation is evaluated via a statistical method described in Ricard and Royer (1993). This new cloud-precipitation-vertical diffusion scheme follows turbulence properties to estimate the statistical water distribution at each grid point. The penetrative convection has been modified to account for the vertical dependency of the air entrainment process (Terray, 1998) and the closure condition is based on humidity convergence criteria. The convective nebulosity and the liquid water fraction are not vertically uniform any more and thus allow a more realistic representation of cumulonimbus anvils. Some refinements related to the snow fraction calculation and to the water total reservoir have been implemented in the land model.

1.2 The oceanic component

The primitive equation model employed in this study is the OPA8 Ocean General Circulation Model (OGCM) (Madec et al., 1998) with a free surface formulation (Roullet and Madec, 2000) in its global configuration ORCA. The horizontal resolution is 2° in longitude and varies from 0.5° at the equator to 2°cos(φ) poleward from the tropics. In order to remove the North Pole singularity, two poles are introduced in the Northern Hemisphere grid (one over the Canada and the other over Siberia). The departure from a geographic mesh starts at 20°N, and the stretched part of the mesh is constructed following the semi-analytical method of Madec and Imbard (1996). There are 31 vertical levels with the highest resolution (10 m) in the upper 150 meters and the time step is 1h 36mn.

The Jackett and McDougall (1995) equation of state is used. It provides density as a function of potential temperature, salinity and pressure. Vertical mixing coefficients are computed from a 1.5 turbulent closure scheme (Blanke and Delecluse, 1993) that permits an explicit formulation of the mixed layer as well as minimum diffusion in the thermocline. Lateral viscosity is laplacian and along the horizontal, with an eddy coefficient varying from 40000 m^2/s at high latitudes to 2000 m^2/s in the equatorial band at the ocean surface, and increasing with depth in the deep ocean. Lateral diffusivity on tracers is along isopycnal surfaces, with a horizontally varying eddy induced coefficient for the Gent and Mc Williams (1990) parameterization of mesoscale eddy induced turbulence. Zero fluxes of heat and salt and no-slip conditions are applied at solid boundaries.

The model includes a sea-ice component, the LLN sea-ice model, developed at the UCL in Louvain-La-Neuve by Fichefet and Morales Maqueda (1997). It runs on the same grid as the ocean model. This model has a representation of both thermodynamic and dynamic processes. A 3-layer model, which takes into account sensible and latent heat storage in the snow-ice system, simulates the changes of snow and ice thickness in response to surface and bottom heat fluxes. The variation of ice compactness due to thermal processes is a function of the energy balance of the surface layer in the region occupied by leads (Fichefet and Morales Maqueda, 1997). To calculate ice dynamics, sea-ice is considered to behave as a viscous-plastic continuum. At the ice-ocean interface, the sensible heat flux is proportional to the temperature difference between the surface layer and its freezing point and to the friction. The ice-ocean stress is taken to be a quadratic function of the relative velocity between ice and the uppermost level of the ocean. Considering salt and freshwater exchanges between ice and ocean, brine is released to the ocean when ice is formed, while freshwater is transferred to the ocean when sea-ice or snow melts.

1.3 The experimental design

The ARPEGE and ORCALIM (ORCA/Louvain Ice Model) models are coupled through the OASIS 2.5 coupler developed at CERFACS (Valcke et al., 2001), which ensures the time synchronization between the GCMs and does the spatial interpolation from one grid to another. Daily mean sea surface temperature (SST), sea-ice extent and surface albedo are given to the atmosphere GCM, and daily mean surface fluxes of heat, momentum and fresh water (including runoff and calving) are given to the ocean GCM. Global flux conservation is performed when passing the heat and water fluxes. All concentrations of greenhouse gases are fixed at pre-industrial values.

The atmospheric initial state is in January of an uncoupled integration of ARPEGE. The initial ocean state is taken from Levitus (1998) for temperature and salinity, and is zero for the velocity field. The initial sea-ice state is also from Levitus (1998).

From this initial state, the coupled model is integrated for one year. During the first year, an excess of fresh water release occurs in the Labrador Sea, altering the deep convection of water masses in this region. To correct this bias, the second year of the integration starts with Levitus (1998) conditions on temperature and salinity fields. Hence, the excess of freshwater is removed from the ocean, enhancing deep convection and limiting sea-ice formation in the Arctic region at a more reasonable value. From this point, the integration is pursued over 69 years to allow the system to reach a quasi equilibrium state in the upper layers of the ocean. We then integrate the coupled model for 200 years. We only consider the latest 200-year period for the subsequent analyses.

1.4 Data storage

The output from the coupled model experiment is available on the Météo-France storage machine “delage”, in the directory: /exterieurs/cglo/cglo315/PR1/sortie, or on the CERFACS “imhotep” machine, in the directory: /home/globc/varia/sim/PR1. More information about the simulation is available at the following address:

http://www.cerfacs.fr/~maisonna/Predicate/couplage.html

2 Overview of the simulation main features in the North Atlantic region

2.1 SST mean state: comparison to Levitus (1998) climatology

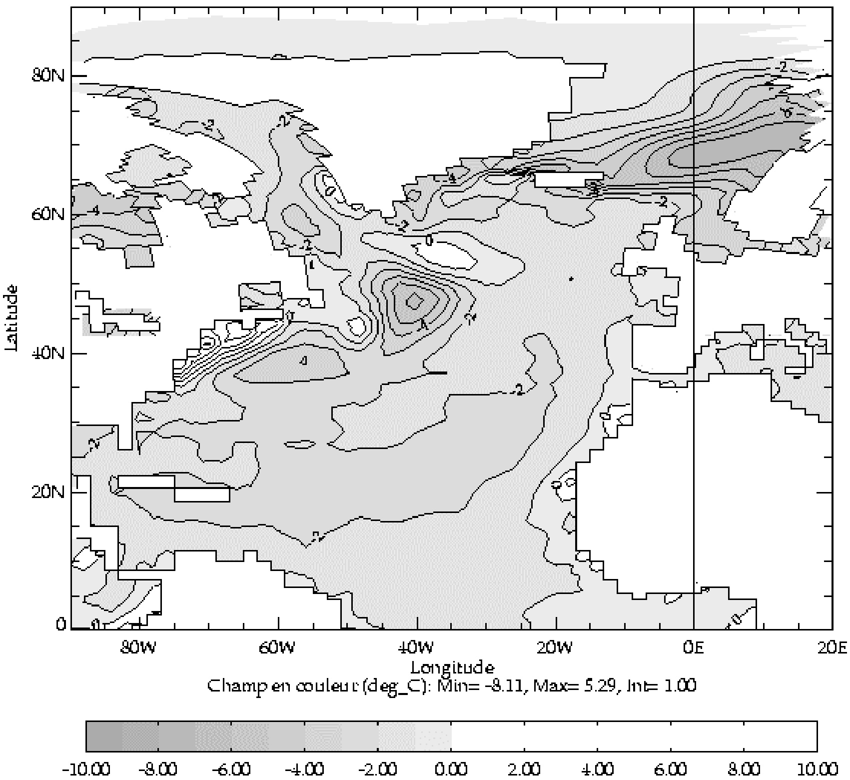


Figure 1: Difference between the coupled simulation SST averaged over the 200 years of integration and Levitus (1998) climatology.

The model SST shows a broad cooling in the whole North Atlantic region compared to Levitus (1998), except off the North American north of Cape Hatteras where positive differences reach 5.3 degrees and on two spots along the North African coast where differences up to +1 degree occur. The maximum cooling occurs in the North Atlantic Current (NAC) region where it reaches –6 degrees, and in the Greenland Iceland Norwegian (GIN) Sea where the differences reach –7 degrees. The main reason for this cooling in the GIN Sea is that the ice extent is much greater in the model than in the climatology (see Appendix). In the subtropics, the differences are small, at about –2 degrees.

2.2 Variability in the North Atlantic region

2.2.1 Winter Mean Sea Level Pressure (MSLP): comparison to NCEP reanalysis

A substantial portion of the climate variability over the Atlantic basin is associated with the NAO, which is a dominant pattern of atmospheric circulation variability. The NAO refers to a meridional oscillation in atmospheric mass with centers of action near Iceland and over the subtropical Atlantic from the Azores across the Iberian Peninsula. When the NAO is in its positive phase, low-pressure anomalies over the Icelandic region and throughout the Arctic combine with high-pressure anomalies across the subtropical Atlantic to produce stronger-than-average westerlies across middle latitudes. This phase of the oscillation is, consequently, associated with cold conditions over the northwest Atlantic and warm weather over Europe, as well as wet conditions from Iceland through Scandinavia and dry conditions over southern Europe. This pattern of climate anomalies is most pronounced during winter when the baroclinic activity is strongest. Considering this influence of the NAO on wintertime European climate, it is of crucial importance that this phenomenon be well simulated in model experiments.

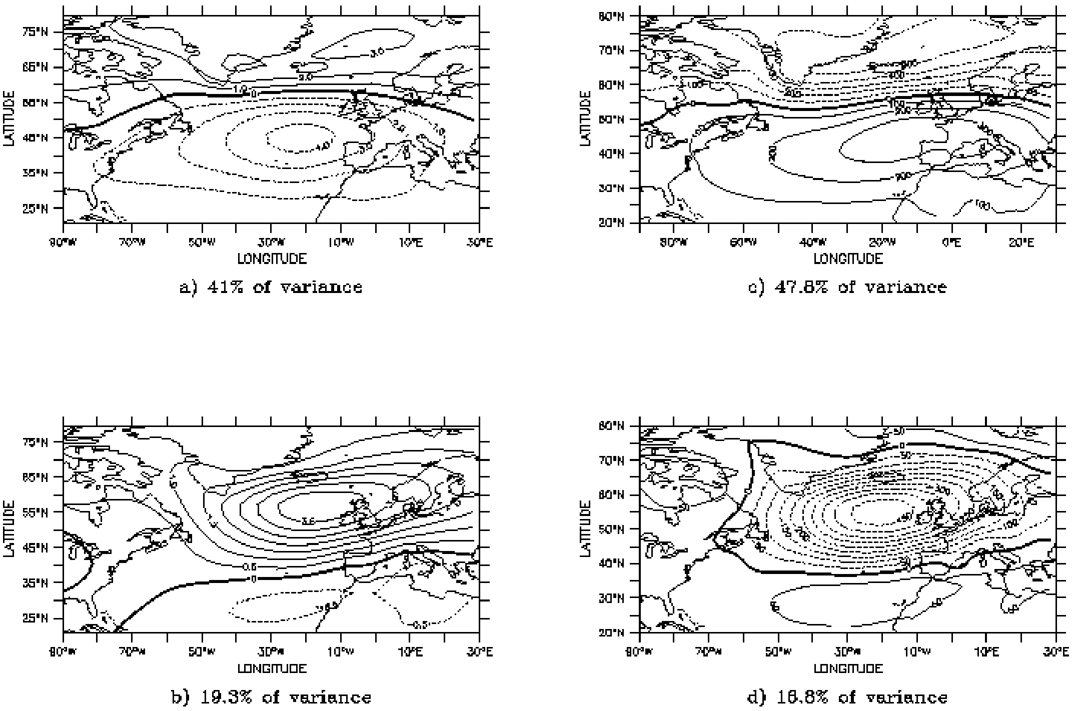


Figure 2: First two EOF modes of December-January-February (DJF) MSLP for the model (a, b, units in hPa) and NCEP reanalysis (c, d, units in Pa). The analysis is performed on the 200 years of the model simulation and on the period 1949-1999 for NCEP.

The first EOF mode of MSLP for the model (Figure 2a) and for NCEP (Figure 2c) represents the NAO mode. The two patterns are in overall agreement, with a more zonal structure for the model. Moreover, the explained variances are of the same order of magnitude. The second EOF mode for the model (Figure 2b) and for NCEP (Figure 2d) depicts the East Atlantic mode, defined by a single pole west of the British Islands. Again, the two modes are similar in shape and explained variances.

Figure 3 illustrates the good agreement between the NAO index as calculated by Hurrell (1995) and the principal component of the first EOF mode, both computed using the model data. The correlation between the two indexes reaches –90 %.

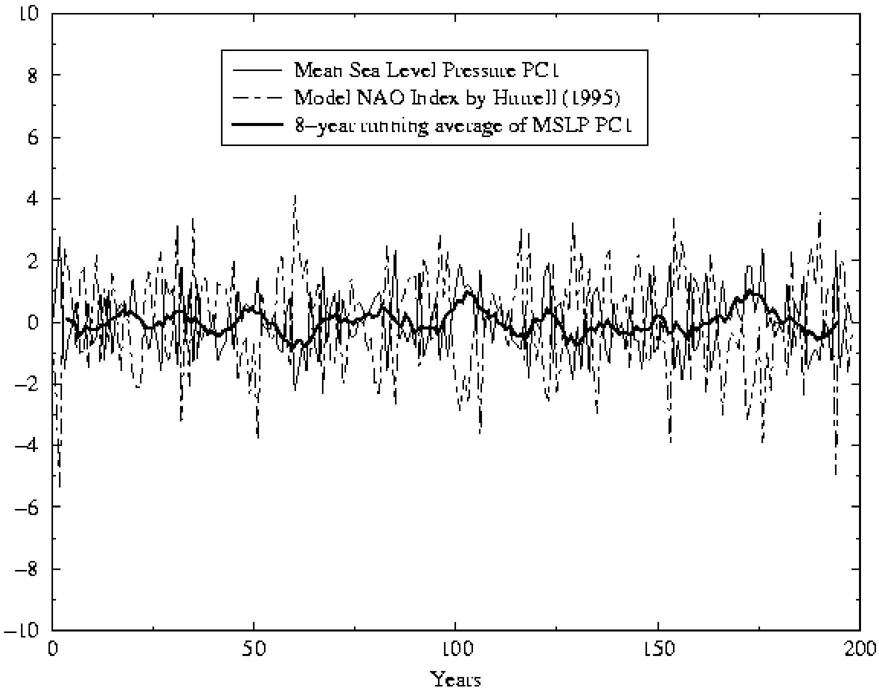


Figure 3: DJF NAO Index by Hurrell (dashed dotted line) versus DJF MSLP first principal component (solid line) and 8-year running average (bold solid line) for the 200-year simulation.

Figure 4 and 5 display the wavelet analysis for the simulation and NCEP respectively. Examination of the global wavelets (Figure 4c, 5c) shows that the model's PC has a nearly flat spectrum (characteristic of a white noise process) with an intermittent peak in the 2-5 year band and a persistent but not significant peak in the 15-30 year band. Conversely, the NCEP PC spectrum is slightly redder with an intermittent peak in the 2-4 year band and a more persistent one in the 6-10 year band.

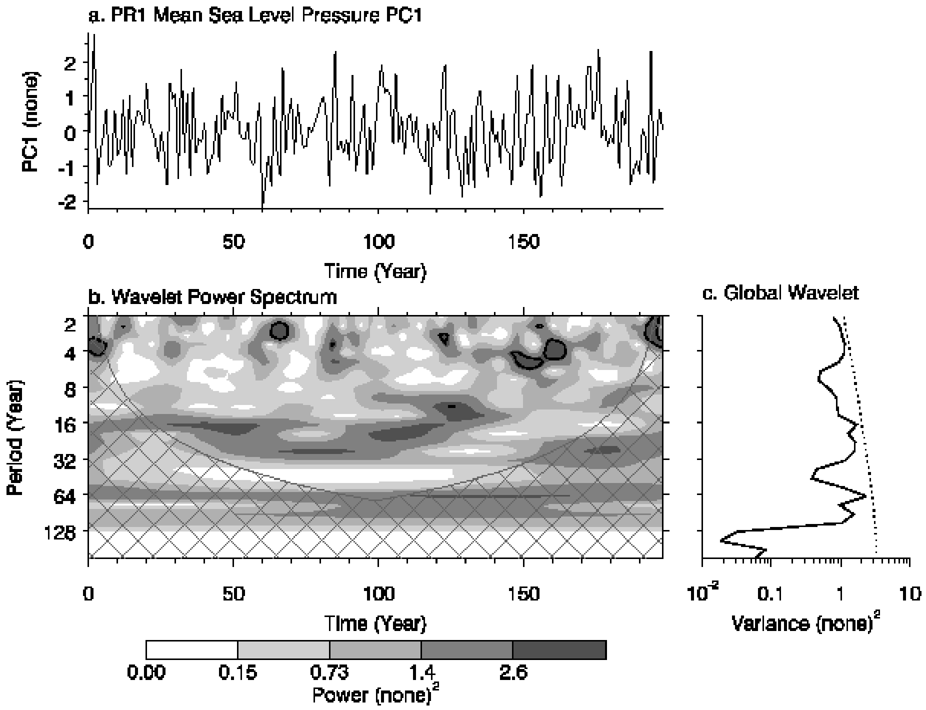


Figure 4: (a) Coupled model DJF MSLP PC1 time series; (b) Wavelet power spectrum of MSLP; solid black line indicates the 95 % confidence level, solid grey line the cone of influence; (c) Global wavelet analysis; the dotted line indicates the 95 % confidence level.

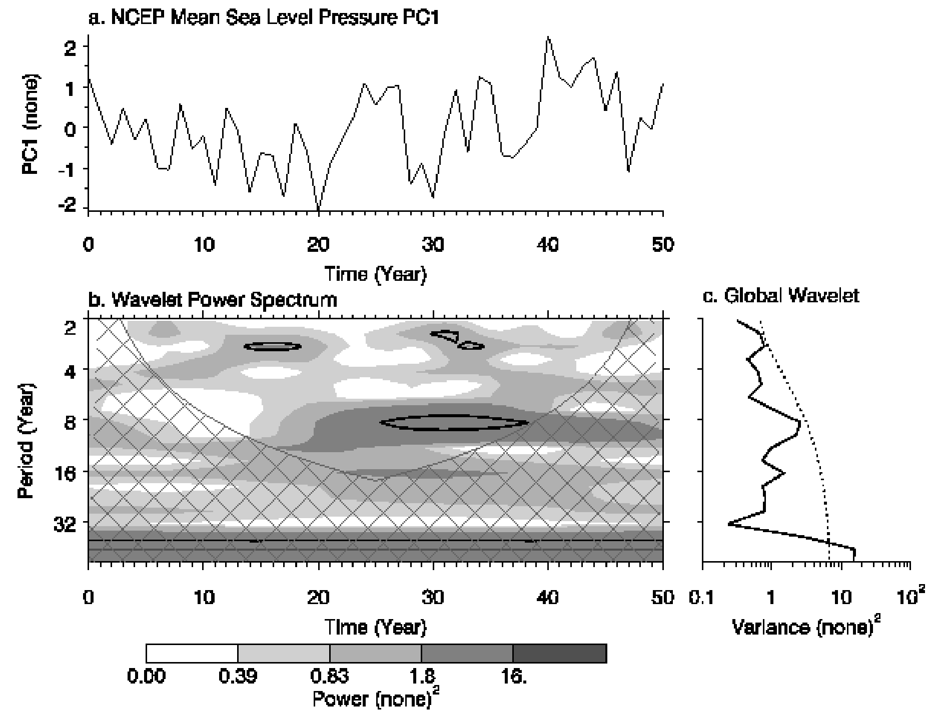


Figure 5: As in Figure 4, but for NCEP MSLP PC1.

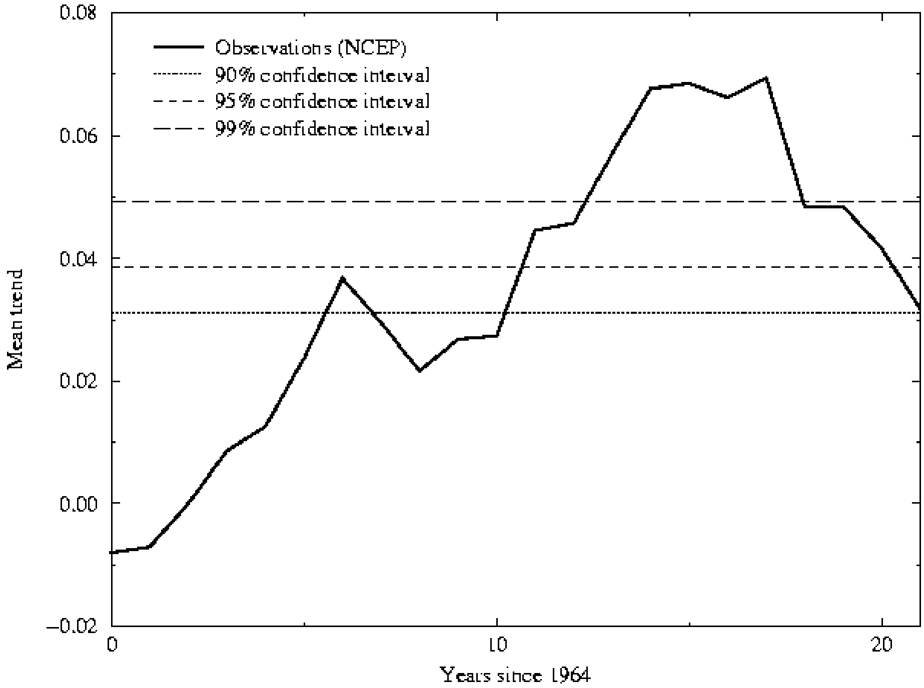


Figure 6: Observations' MSLP PC1 mean trend calculated using a 30-year running window. Also shown are the 90 %, 95 % and 99 % confidence intervals for the model's mean trend.

Figure 6 exhibits the mean trend in NCEP MSLP PC1. After a steady, nearly constant, increase during the first 15 years, the trend decreases during the last 5 years. Compared to the model's confidence intervals, the observations show a peak during the 1980s, which is not reproduced by the model. This means that the model underestimates reality or that the observations include a component which is not present in the model (for example, the human-induced increase in greenhouse gases concentrations).

2.2.2 The Thermohaline Circulation (THC) in the model

The thermohaline circulation (THC) in the North Atlantic Ocean plays an essential role in the maintenance of the current climate. Warm, salty water from tropical and subtropical latitudes is transported northward to relatively high latitudes. The cold near-surface atmosphere at these latitudes is very effective at extracting heat from the ocean, thereby allowing the water to cool, increase in density, and sink. The water then flows equatorward at depth as North Atlantic Deep Water. This process contributes to the total oceanic poleward heat transport.

As the atmosphere has a very short memory (on the order of one month), any interannual to decadal predictability in the coupled system can only arise from an influence of the oceanic state on atmospheric parameters. There are recent indications that variations of the dominant SST patterns in the North Atlantic which have been associated with climate changes over Europe and Eurasia can indeed be predicted over a decade or longer, provided that the state of the THC is monitored with adequate accuracy (Griffies and Bryan, 1997). Therefore, it is of crucial importance to understand the processes causing variations in the THC and to assess the potential predictability of such fluctuations.

The coupled model used here will allow us to investigate the processes underlying the variability of the THC at decadal timescales. To determine the potential predictability of fluctuations in the THC, two 25-year long ensemble experiments, using the same model, have been carried out. For each ensemble, the different members (6 members for each ensemble) only differ by infinitesimal perturbations of their initial atmospheric conditions. On the other hand, the different ensembles contrast different initial oceanic conditions. The first ensemble starts at a maximum of the intensity of the THC, the second one at a minimum. Hence, it will be possible to study the sensitivity of the predictability to the phase of the THC.

Figure 7 shows the two first EOFs of the North Atlantic meridional streamfunction (hereafter MSF) and their associated PCs. The first EOF depicts a basin-wide cell with a maximum between 30°N and 50°N at a depth of about 2000 meters. The PC of this mode shows low frequency fluctuations at a period between 30 and 50 years, significant at the 95% confidence level (Figure 8). The second EOF mode depicts two cells of opposite sign, extending from the surface to the bottom of the ocean, resembling the Ekman response to the surface wind forcing of the ocean. The separation between the two cells occurs at 35°N. The PC of this mode has a nearly white spectrum with significant fluctuations at the decadal time scale (Figure 9).

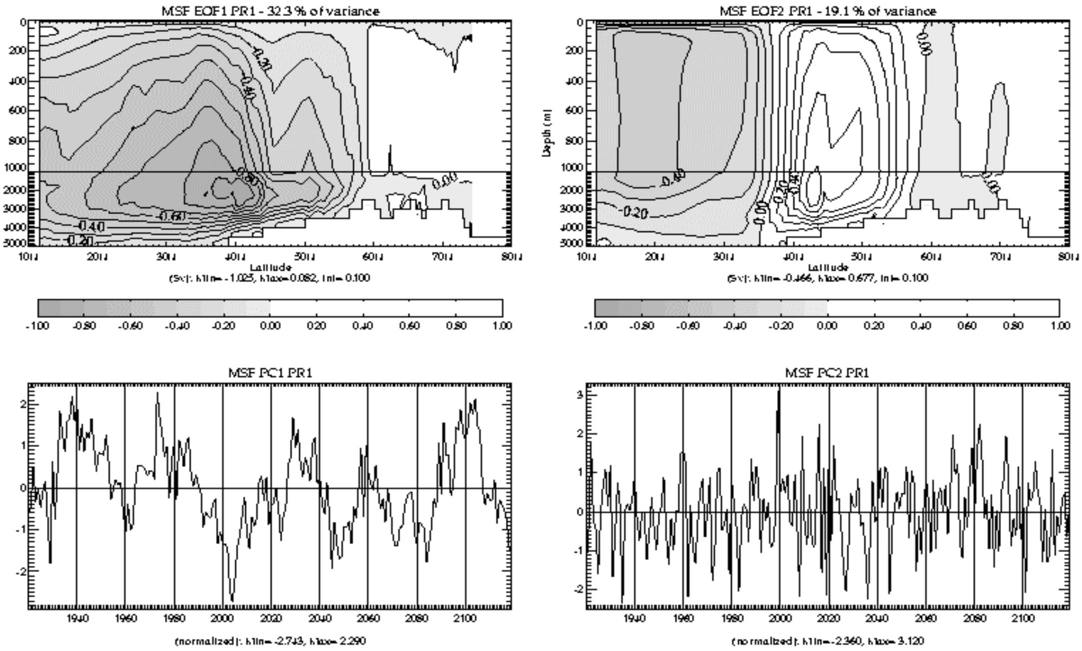


Figure 7: First Two EOF modes of variability and their associated PCs for the North Atlantic MSF.

(a) (b)

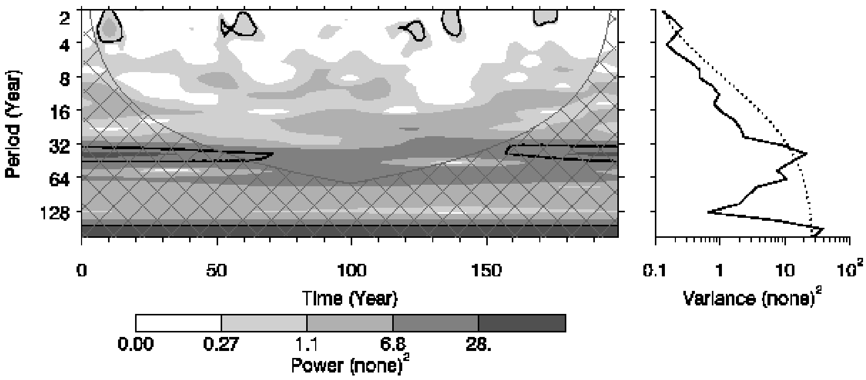


Figure 8: (a) Wavelet power spectrum for the MSF PC1 time series; solid black line indicates the 95% confidence level, solid grey line the cone of influence; (b) Global wavelet analysis; the dotted line indicates the 95% confidence level.

(a) (b)

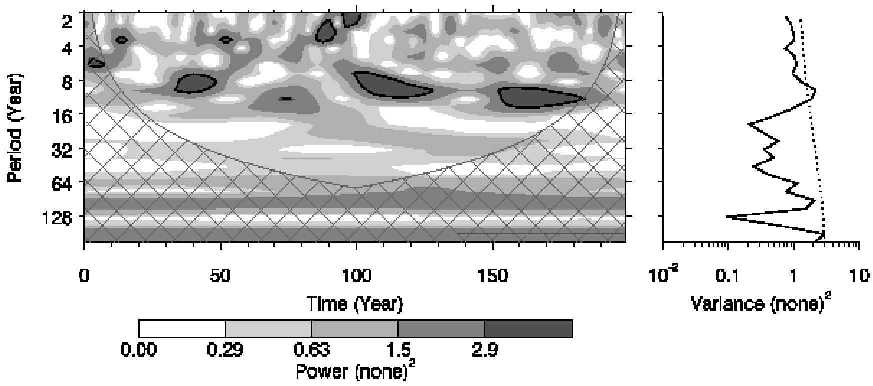


Figure 9: Same as Figure 8 but for MSF PC2 time series.

The THC index is defined as the maximum of the Atlantic MSF north of 20°N. It gives a measure of the intensity of the overturning circulation in the Atlantic basin. Figure 10 shows a comparison between this index filtered to remove periods lower than 20 years and the first PC of the North Atlantic MSF filtered in the same frequency band. The two series are highly correlated (-0.98), meaning that the first PC of the North Atlantic MSF gives a good measure of the low frequency fluctuations of the THC.

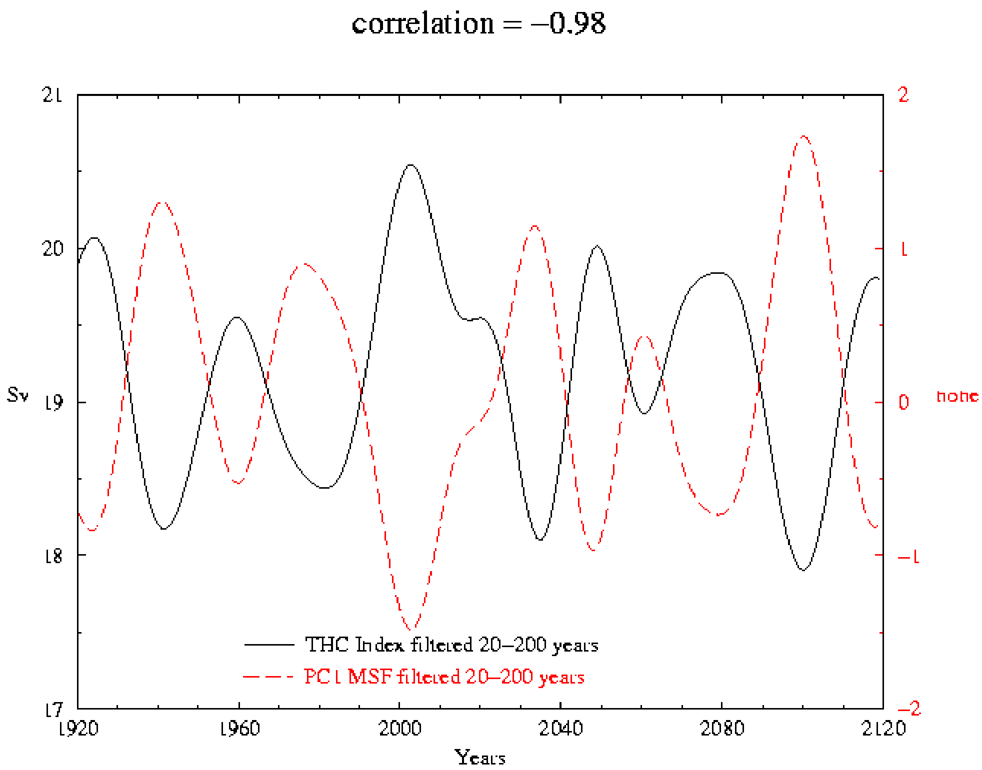


Figure 10: THC index and PC1 of North Atlantic MSF both filtered in the band 20-200 years.

2.2.3 Relations between low frequency fluctuations of the THC and various fields

In the precedent section, we've seen that there were two low frequency modes of variability: the first one in the band 30-50 years, the second one in the band 8-12 years. The second mode is linked to fluctuations in the surface wind forcing of the ocean, while the first mode describes multi-decadal fluctuations of the overturning cell. Figure 11 depicts the lagged correlations between the MSF PC2 and the NAO index. The maximum correlation is obtained at lag 0. The regression at this lag between the MSF PC2 index and the SST and zonal wind is shown in Figure 12. The picture obtained is the well-known tripole pattern of SST and the NAO-like pattern in zonal wind. Thus, this mode is excited by rapid fluctuations in the NAO.

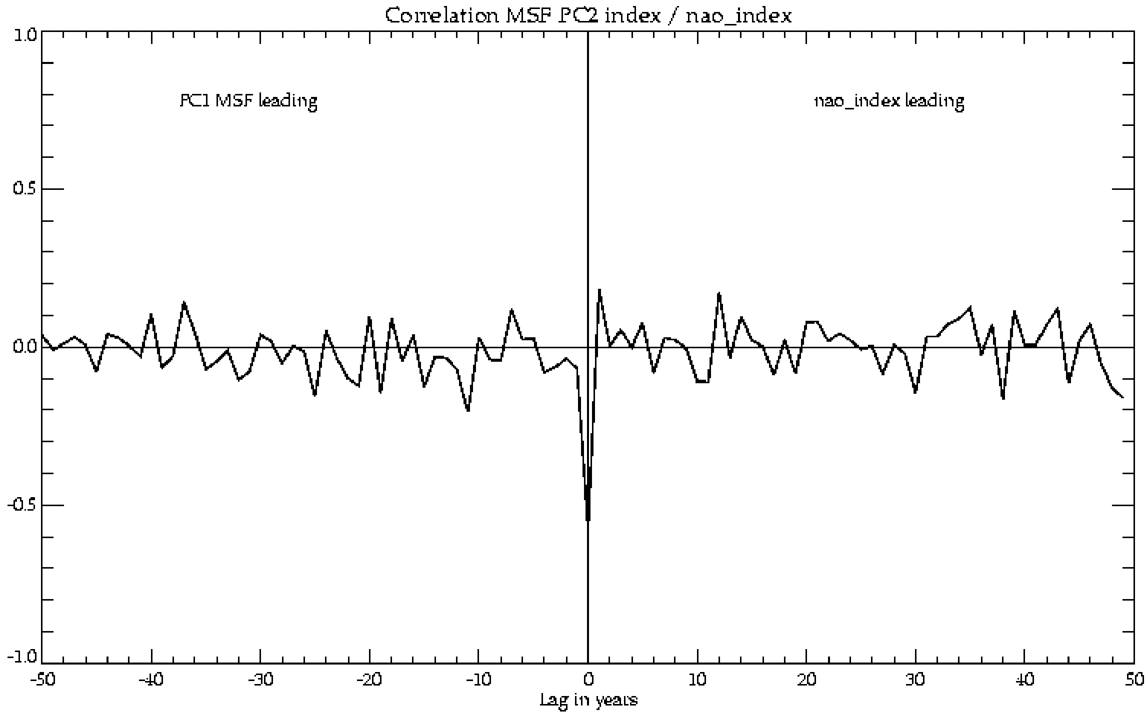


Figure 11: Correlation between the North Atlantic MSF PC2 and the NAO index as calculated by Hurrell (1995). Positive lags indicate that the NAO index is leading.

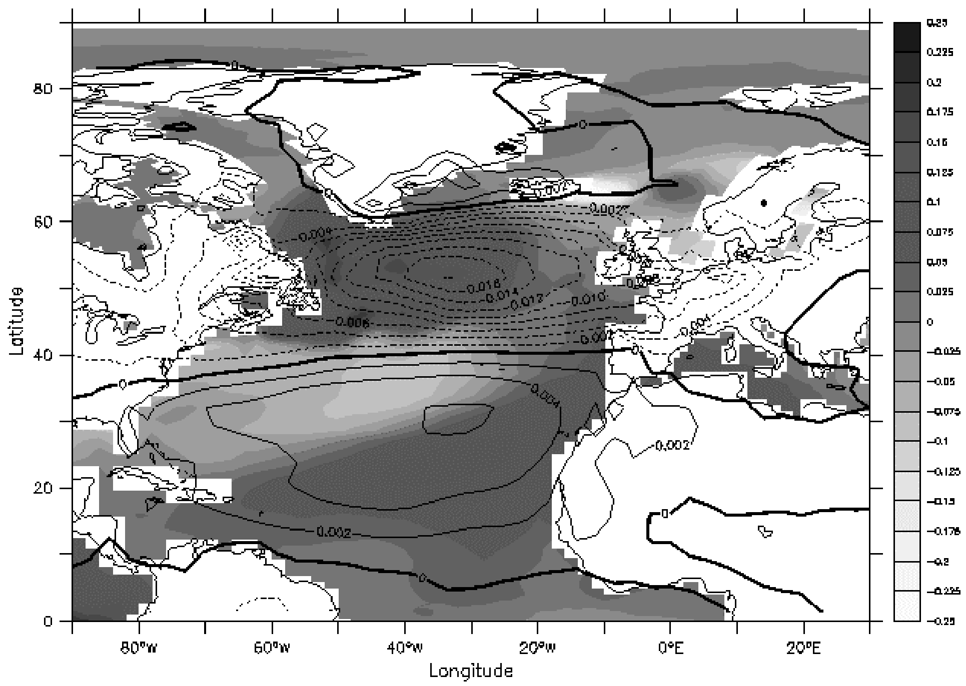


Figure 12: Regression at lag 0 between the North Atlantic MSF PC2 and SST (shading) and zonal wind (contour). Shading interval is 0.025 °C, contour interval is 0.002 N/m^2.

The reason for the existence of the multi-decadal mode is less clear. The correlation between the MSF PC1 and the NAO index remains very weak at all lags (not shown). The signature of this mode in terms of SST and Sea Surface Salinity (SSS) is shown in Figure 13.

(a) (b)

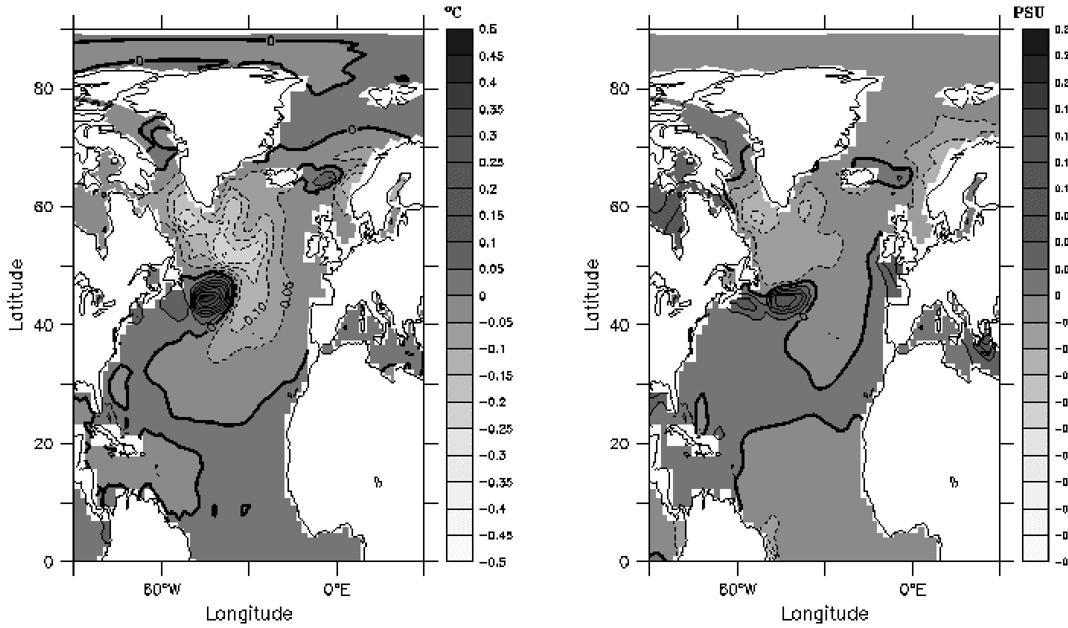
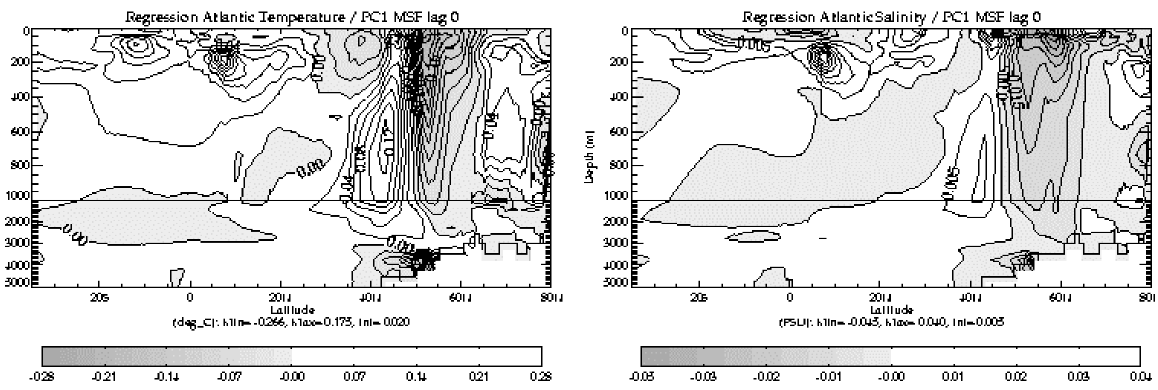


Figure 13: Lag 0 regression between the MSF PC1 filtered in the band 20-200 years and (a) SST, (b) SSS. Contour interval is 0.05 °C for (a) and 0.025 PSU for (b).

For both SST and SSS, the patterns emerging are similar. It consists in a warm, salty region off Newfoundland surrounded by a cold and fresh tongue extending from 30°N to high latitudes. This picture is consistent, in terms of SST, with a weaker northward heat transport resulting from a weaker THC (a high value of the MSF PC1 corresponds to a low value of the THC index, see Figure 10).



(a) (b)

Figure 14: Lag 0 regression between the MSF PC1 index filtered in the band 20-200 years and (a) the Atlantic zonal mean temperature, and (b) the Atlantic zonal mean salinity. Contour interval is 0.02 °C for (a) and 0.005 PSU for (b).

Figure 14 depicts the lag 0 regression between the MSF PC1 filtered in the band 20-200 years and Atlantic zonal means of temperature and salinity. It appears that the anomalies seen in Figure 13 are not surface trapped but extend from the surface to the bottom of the ocean. Hence, there must be some kind of oceanic mechanism that provokes the existence of such anomalies.