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Statement on the changes in natural modes of climate variability under anthropogenic climate change based on scenario and time-slice simulations

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ENSEMBLES
Deliverable D4.2.4

WP number: 4.2

Participants: CERFACS, CNRM, IfM-Kiel, INGV, IPSL, UREADMM

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Outline:

1. Impacts of climate change on ENSO variability
2. Impacts of climate change on Monsoons and their teleconnections
3. Change in extratropical variability modes: the Euro-Atlantic sector
Abstract

The primary objective of RT4 is to advance understanding of the basic science issues at the heart of the ENSEMBLES project. This aim is pursued through the elucidation of the key processes that govern climate variability and change, and that determine the predictability of climate on timescales of seasons, decades and beyond. One of the task of the research activity is to determine the impact of climate change on climate variability. Within WP4.2, research activity has been carried out to explore the characteristics of global and regional climate variability modes and how these might be affected by global warming, using both model and observational data sets.

The research, which has been performed in close collaboration with RT5, has examined the major modes of variability on timescales from seasons to decades, for different regions of the globe, with an emphasis on scale interactions and modes interaction. For the tropical region, for example, the major influences of the global warming on the most important modes of variability, such as the Asian monsoon and ENSO, and on their interaction have been investigated and quantified. The analysis has produced an extensive and detailed characterization of how the natural variability might be affected by global change and the results have been published in numerous scientific papers (see references).

In the following, the main results obtained from the partners who have contributed to this characterization are summarized. These results represent a useful reference for the evaluation of the ENSEMBLES models, as well as a control case for the assessment of the possible impacts of climate change on the characteristics of the climate variability.
1. Impacts of climate change on ENSO variability

1.1 Changes of ENSO variability in IPCC AR4 models

A review on “Understanding El Niño in Ocean-Atmosphere General Circulation Models: progress and challenges” has been published in the *Bull. Amer. Met. Soc.* (Guilyardi et al. 2009) and the part relevant to ENSO in a changing climate is given below.

Most (but not all) IPCC AR4 models are qualitatively consistent in their projections of mean changes over the tropical Pacific. The SST warms more along the equator than off the equator, and a reduced east-west SST gradient is associated with a weakened Walker circulation and reduced trade winds (Hansen et al. 2006, Fedorov et al. 2006, Vecchi et al. 2006, 2008). Such changes in the mean state can influence the ENSO-related processes and feedbacks and have the potential to modify ENSO properties. For example, studies show that a more stable ENSO is less sensitive to changes in the background state than when it is closer to instability (Zelle et al., 2005). Atmosphere deep convection triggering is also highly dependent on the mean SST distribution, and associated heat flux feedbacks may change. Nevertheless, van Oldenborgh et al. (2005) noted that if only the six “best” models for ENSO are considered, the tendency for reduced mean east-west gradient is much less obvious than if all models are considered.

However, and as seen from Fig. 1.1.1 (Fig. 10.16 of the IPCC AR4 report), which displays the ratio of ENSO variability between the current climate and the last 50 years of the SRES A2 experiments (2051–2100) as a function of the background change, models are inconsistent with respect to their projections of change in ENSO amplitude (see also van Oldenborgh et al. 2005, Merryfield 2006, Guilyardi 2006) even in very high CO2 scenarios (Fig. 5 in Guilyardi et al., 2009). While some models show an increase in ENSO variability in response to greenhouse gas increases, others do not exhibit any detectable change while still others show a decrease in variability.


Nevertheless, changes of ENSO variability, where they can be detected above these large natural variations, are highly model dependent, even if extreme scenarios are analyzed (4xCO2). Hence, even though all models show continued ENSO variability in the future no matter what the change of average background conditions, there is no consistent indication at this time of discernible changes in amplitude or frequency for the 21st century (Meehl et al. 2007b). Similarly, large model differences in the skewness of the variability limit the assessment of the future relative strength of El Niño and La Niña events (van Oldenborgh et al. 2005). Because ENSO is the dominant mode of climate variability at interannual timescales, the lack of consistency in the model predictions of the response of ENSO to global warming currently limits our confidence in using these predictions to address adaptive societal concerns, such as regional impacts or extremes (Joseph and Nigam 2006, Power et al. 2006). Nevertheless, paleo-evidence that ENSO may have been quite different in the past (e.g. Tudhope et al. 2001, Cobb et al. 2003) indicates that there is a risk ENSO and the associated teleconnections (see for instance Meehl and Teng 2007 on the shift of ENSO teleconnections in North America) might be quite different in the future: a fact that is available to those assessing mitigation options.

A better understanding of the sensitivity of ENSO to changes in processes and feedbacks will help explain these differences, possibly leading to more confident
projections. For instance, the disagreement among the various IPCC AR4 models regarding future changes in ENSO does not rule out that a subset of models can show a common ENSO response to climate change. Guilyardi (2006) showed that among those models that best reproduced the diversity of the observed ENSO, there was a significant trend towards increased El Niño amplitude in high CO2 scenarios. Hence, to improve decadal to centennial projections, process and feedback diagnostics are needed to limit the subset of models to those that are more consistent with the real world. Even if models do not predict significant changes in El Niño statistics in the future (e.g. amplitude or frequency), the relative balance of feedbacks and teleconnections (and the associated impacts) during ENSO could evolve (Philip and van Oldenborgh, 2006), perhaps altering ENSO predictability.

**Fig. 1.1.1.** Mean state change in average tropical Pacific SSTs and change in El Niño variability simulated by AOGCMs. The mean state change (horizontal axis) is computed over the area 10S to 10N, 120E to 80W (reproduced from Yamaguchi and Noda 2006). The change in El Niño variability (vertical axis) is denoted by the ratio of the ENSO amplitude between the current climate and the last 50 yr of SRES A2 experiments (2051-2100), except for FGOALS-g1.0 and MIROC3.2(hires), for which the SRES A1B was used, and UKMO-HadGEM1, for which the 1% yr\(^{-1}\) CO2 increase climate change experiment was used, in the region 30S to 30N, 30E to 60W (reproduced from van Oldenborgh et al. 2005). Error bars indicate the 95% confidence interval [Figure reproduced from Guilyardi et al. (2009)].

1.2 ENSO response to global warming: review and new simulations with KCM model
In an effort to further understand ENSO response to global warming, an ensemble of eight greenhouse warming simulations was performed with the Kiel Climate Model (KCM), in which the CO$_2$ concentration was increased by 1% yr$^{-1}$ until doubling was reached, and stabilized thereafter (Park et al. 2009). Multi-century integrations with KCM indicate the model simulates tropical Pacific climate well; in particular, the model simulates a relatively small cold bias (1°C), an annual cycle matching observations in amplitude and phase, and the amplitude and period of interannual variability in agreement with observations. Warming of equatorial Pacific sea surface temperature (SST) under CO$_2$ doubling is, to first order, zonally symmetric and leads to a sharpening of the thermocline. ENSO variability increases because of global warming: during the 30-yr period after CO$_2$ doubling, the ensemble mean standard deviation of Niño-3 SST anomalies is increased by 26% relative to the control, and power in the ENSO band is almost doubled. The increased variability is due to both a strengthened (22%) thermocline feedback and an enhanced (52%) atmospheric sensitivity to SST; both are associated with changes in the basic state. Although variability increases in the mean, there is a large spread among ensemble members and hence a finite probability that in the “model world” no change in ENSO would be observed (Figure 1.2.1).

**Figure 1.2.1**: SST anomalies averaged over the Niño3 region ($150^\circ$W–90$^\circ$W, 5$^\circ$N–5$^\circ$S) in an ensemble of eight 1% CO$_2$ integrations with the Kiel Climate Model (KCM). Ensemble spread (gray shading), ensemble mean, a member with increasing variability, and one with no change in variability are shown. [Figure reproduced from Latif and Keenlyside (2008)].

As reviewed above, and also by Latif and Keenlyside (2008), there is no clear evidence for significant tipping point behaviour with respect to ENSO variability. However, although there may be no tipping point in the physical system, it does not preclude the existence of tipping points in other components of the Earth System induced by changes in the Tropical Pacific climate. For example, Cox et al. (2004) show, in a physical climate model coupled to a carbon cycle component that includes dynamic vegetation, that a slow
change toward an El Niño-like state leads to a smooth reduction in precipitation in the Amazon (Fig. 1.2.2a). This produces a rapid collapse of the rainforest in the mid-twenty-first century, and results in the Amazon becoming a source of atmospheric CO$_2$, instead of sink as in present-day climate. These changes contribute significantly to the global mean uptake of carbon by land, which also becomes a carbon source at 2050 (Fig. 1.2.2b). This leads to a significant feedback in the carbon cycle in which the airborne fraction of CO$_2$ increases nonlinearly, accelerating global warming. This is just one example, and other tipping points may be lurking in biological, chemical, or even socioeconomic systems. Much research is required to investigate such possibilities and their physical drivers and interactions. The rate of physical-system change, even if it is smooth, may also be an issue for the criticality of other systems. This highlights the importance of including new components into Earth System models, as has been done in ENSEMBLES.

Figure 1.2.2: ENSO and carbon cycle feedbacks on climate involving the Amazon rain forest. (A) Relationship between the Equatorial Pacific zonal SST gradient and Amazon rainfall for winter as observed (asterisks) and simulated by the Hadley Centre model (diamonds), which includes climate feedbacks on the carbon cycle. Linear relations are indicated for the model (present and future) and observations. In the model, the SST gradient weakens in the future, leading to reduced Amazonian rainfall and rainforest dieback. (B) Global land carbon uptake is sensitive to climate change. When climate feedbacks on the carbon cycle are included (solid) land becomes a source of atmospheric carbon around 2050; Amazon rainforest dieback is a major contributor. When they are not included (dashed) land always remains carbon sink. [Figure reproduced from Latif and Keenlyside (2008)].
2. Impacts of climate change on Monsoons and their tele-connections.

2.1 Influence of snow cover on the Indian Summer monsoon variability

The possible influence of the winter/spring Eurasian snow cover on the subsequent Indian summer precipitation has been revisited using both observations and a subset of CMIP3 simulations (Peings and Douville 2009). In keeping with former studies, the observations suggest a link between an east-west snow dipole over Eurasia and the Indian summer monsoon precipitation. However, this relationship is neither statistically significant nor stationary over the last forty years. Moreover, the strongest signal appears over eastern Eurasia and is not consistent with the Blanford hypothesis whereby more snow should lead to a weaker monsoon. The 20th century CMIP3 simulations provide longer timeseries to look for robust snow-monsoon relationships. The maximum covariance analysis indicates that some models do show an apparent influence of the Eurasian snow cover on the Indian summer monsoon precipitation, but the patterns are not the same as in the observations. Moreover, the apparent snow-monsoon relationship generally denotes a too strong El Niño-Southern Oscillation teleconnection with both winter snow cover and summer monsoon rainfall rather than a direct influence of the Eurasian snow cover on the Indian monsoon.

2.2 Response of Asian monsoon to increased CO2

The Asian monsoon remains robust at 2xCO2, with slight increases in JJAS mean rainfall compared to picture (Fig. 2.2.1). Interannual variability increases however (measured by rainfall or dynamically), related to increased ENSO variability in the equatorial Pacific. In a parallel pair of integrations in which systematic SST biases are corrected in the equatorial Indo-Pacific using heat flux corrections (mainly correcting too warm Maritime Continent and cold bias in the cold tongue) the response to 2xCO2 is found to be larger, suggesting model bias may be masking the true impact of climate change. The monsoon-ENSO teleconnection, important for seasonal prediction, is found to be relatively robust (both dynamically and in terms of precipitation) in the future climate: in other words it does not show the recent weakening seen in observations since the late-1970s. However, the strength of the monsoon-ENSO teleconnection is found to vary widely over the course of all 100-year integrations (in the absence of changes to external forcing, Fig 2.2.2) suggesting that internal variations may be important (Turner et al., 2007a).
2.3 Monsoon-ENSO tele-connections

We demonstrate a future scenario in HadCM3 in which ENSO undergoes regime-like behaviour, each lasting several decades (Fig 2.3.1). In a periodic regime, ENSO is found to be strictly biennial and strongly coupled to the Asian-Australian monsoon via the TBO. In an irregular ENSO regime, the monsoon-ENSO teleconnection is much weaker (Fig 2.3.2). Should such behaviour occur in reality, it would have strong implications for the monsoon systems (Turner et al., 2007b).
2.4 Effects of increased CO2 concentrations on monsoon intraseasonal variability.

The monsoon intraseasonal variability under 2XCO2 conditions is investigated. We show that HadCM3 can simulate northward propagating ISO/active-break cycles. These events are shown to be more intense at 2xCO2 (breaks when measured against the wetter annual cycle, Fig. 2.4.1), although there is no suggestion of any change to their timing or duration. Monsoon extremes are found to intensify also, consistent with other work on tropical rainfall. The increases in (strength of) heaviest monsoon extremes in this model are found to intensify at a rate predictable by local surface warming and Clausius-Clapeyron (Fig 2.4.2). A further paper examines this relationship in the CMIP3 models and finds that 6 of 15 studies also share this behaviour. 3/15 models show weakened or no change to the heaviest extremes, while a further 6/15 show increases far beyond thermodynamic predictions. There is some suggestion that the type of convective parameterization may play a role in this behaviour (Table 2.4.1) [Turner et al (2009a,b)].
Figure 2.4.1: Differences between lag-zero composites of active (left) and break (right) events in the Indian summer monsoon at 1xCO2 and 2xCO2. Upper panels show absolute differences. Lower panels show differences after removing the change in mean seasonal cycle [Figure reproduced from Turner et al. (2009a)].

Figure 2.4.2: Levels of precipitation in the upper quartiles for the Indian region in HadCM3 1xCO2 (black), 2xCO2 (red) and their ratio (green, on right-hand scale). No spatial averaging
is performed. Asterisk indicates the predicted increase in the heaviest rainfall via the degree of local surface warming and Clausius-Clapeyron [Figure reproduced from Turner et al. (2009a)]

<table>
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<tr>
<th>Model</th>
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<th>Predicted (%)</th>
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<th>Convection</th>
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Table 2.4.1: List of CMIP3 models showing measured and predicted increases in intensity of the heaviest rainfall, ranked by their ratio. Models are split into 4 categories: super-thermodynamic (top), thermodynamic (middle) and near-zero (bottom). Colouring indicates models with identical convection schemes. AS indicates Arakawa-Schubert [Figure reproduced from Turner et al. (2009b)].

3. Change in extratropical variability modes: the Euro-Atlantic sector

3.1. Changes of NAO variability in response to increase of greenhouse gas forcing in the time slice experiments

Observations show that there was change in interannual North Atlantic Oscillation (NAO) variability in later 1970s. This change was characterized by an eastward shift of the NAO action centres, a poleward shift of zonal wind anomalies and a downstream extension of climate anomalies associated with the NAO. The NAO interannual variability for the period after the 1970s has an annular mode structure and that penetrates deeply into the lower stratosphere, indicating a strengthened relationship between the NAO and the Arctic Oscillation (AO). In this study we have investigated possible causes of these changes in the NAO by carrying out experiments with an atmospheric GCM. The model is forced either by doubling CO2, or increasing sea surface temperatures (SST), or both. Results indicate that SST and CO2 change both force an eastward shift in the pattern of interannual NAO variability and the associated poleward shift of zonal wind anomalies, similar to that seen in observations. The effect of SST change can be understood in terms of mean changes in the troposphere, especially Eady growth rate. The effect of CO2 change, however, can not be understood in terms of mean changes in the troposphere. However, the significant response in the stratosphere is characterized by strengthened climatological polar vortex and enhanced interannual variability. These results could suggest an important role for the stratosphere in explaining the recent eastward shift in the pattern of interannual NAO variability and its related climate anomalies.
3.2. Effects of anthropogenic climate change on weather regimes in CMIP3 21st Century simulations

A paradigm for climate change states that external forcing (for instance anthropogenic forcing) can modify the frequency of occurrence of intrinsic modes of variability without changing drastically their structure or introducing new ones (Palmer 1999, Corti et al. 1999). We have checked whether this paradigm can be applied to the mid-to-high latitude modes of the large-scale atmospheric circulation (LSC) as represented by weather regimes. We have studied changes in the occurrence frequency of weather regimes in the 21st century simulations of the CMIP3 aka ENSEMBLES stream 1) multi-model dataset. The analysis has been carried out for both summer and winter. Here in this short summary report, we focus on the summer season (results for the winter analysis and other LSC variables like the wind at 850 hPa can be found in the referenced papers). Figure 3.2.1 (left) shows the four observed regimes (from NCEP reanalysis) and their links with precipitation (CRU data). The second and fourth regimes can be viewed as the negative and positive phases of the summer North Atlantic Oscillation (NAO), respectively. The NAO+ regime can also be viewed as a blocking-like pattern. The links between regimes and precipitation over Europe is assessed computing the composite of observed precipitation corresponding to each regime (Fig. 3.2.1, right).

![Figure 3.2.1](image.png)

Figure 3.2.1: (left) Composite anomalies of SLP corresponding to the four observed summer regimes (hPa). (right) Relative composite anomalies of precipitation (percentage) corresponding to the four regimes. The frequency of occurrence of the regimes is: 26% (Atl. Ridge), 23% (NAO-), 22% (Atl. Low), 29% (NAO+).

We have studied the occurrence of the regimes in the CMIP3 models. SLP maps from the CMIP3 models are interpolated on NCEP reanalysis grid and then projected on the 10 leading EOF patterns of NCEP SLP. This procedure gives time series that allow defining each day of a CMIP3 model simulation in the space spanned by the 10 leading NCEP EOF. The centroids of NCEP weather regimes are then considered as reference and each day from the
CMIP3 models is classified in a regime by minimization of the distance to the NCEP centroids. This procedure ensures that the reference is the same for all the models in the present and future climate, so that we can compare the occurrence of the regimes for the two periods and among the different models. First, in order to test if the observed clusters are relevant for all models and periods (present and future), two diagnostics are used. In a first time, the mean frequencies of occurrence of the regimes in the present climate from the CMIP3 models are compared to the observed ones (Fig. 3.2.2, left). The very strong similarity in the frequencies of occurrence of the regimes in the models and in the observations gives some indications that the observed regimes are relevant for the CMIP3 models. The second test consists in computing the mean Euclidean distance between the centroid and the SLP pattern of the days that belong to the regime, for each regime and model, both in the present and future climate simulations (Fig. 3.2.2, right). The same distances are also computed for the NCEP. The comparison between the CMIP3 values and the NCEP values provides an indication about how well model days fit in observed clusters. The mean distances to the centroid in the CMIP3 models are generally significantly different than the observed ones, even if the differences are generally small. However, the results of the CMIP3 mult-model in the present climate are very similar to the NCEP ones. It is also interesting to note that the differences between the present and future mean distances to the centroid among the models are very limited: the states obtained in the present climate are still relevant in the future climate, which is coherent with the initial paradigm saying that the anthropogenic climate change may not lead to new regimes but to a change in the residence frequency in the present-day regimes.

**Figure 3.2.2:** (Left) Frequency of occurrence of the weather regimes in the NCEP reanalysis (gray bars) and in the 15 CMIP3 (crosses) in the 1961–2000 period. The black line is the CMIP3 ensemble mean. A black (red) cross indicates that the difference with the observed occurrence is non significant (significant) at the 0.05 level. Right: Mean Euclidean distance between the SLP pattern of the days that belong to the regime and the centroid, as observed (gray bars) and in the present climate simulations (crosses on the left of the bars) and future climate simulations (crosses on the right of the bars). The black (dotted) line is the CMIP3 ensemble mean for present (future) climate. A black (red) cross indicates that the difference with the observed occurrence is non significant (significant) at the 0.05 level. Given the availability of the necessary variables, the models considered in this study are BCC-BCM2.0, CGCM3.1(T47), CGCM3.1(T63), CNRM-CM3, CSIRO-Mk3.0, GFDL-CM2.0, GFDLCM2.1, GISS-AOM, FGOALS-g1.0, INM-CM3.0, IPSLCM4, MIROC3.2(hires), MIROC3.2(medres), ECHO-G, ECHAM5/MPI-OM, MRI-CGCM2.3.2, CCSM3, and PCM.
We now consider the changes in the occurrence of the regimes in response to anthropogenic forcings in the CMIP3 multi-model. Figure 3.2.3 shows the changes in the occurrence of the regimes at the end of the 21st century. In most models, an increase (decrease) of the occurrence of NAO+ and Atl. Ridge (NAO- and Atl. Low) generally occurs. However, the magnitude of these changes strongly varies among the different models. Some models exhibit large LSC changes (CNRM-CM3, GFDL-CM2.1, FGOALS-g1.0), whereas other models (MIROC3.2(medres), MRI-CGCM2.3.2) simulate very limited ones. The increase (decrease) of the occurrence of NAO+ (NAO-) corresponds to a decrease in precipitation over the north of France and United Kingdom and increase over Spain. One can show that the amount of precipitation change for these regions is very strongly correlated to the detailed evolution of these two regimes, in particular the NAO+ or blocking, under anthropogenic forcing (Boé et al. 2008).

![Figure 3.2.3: Changes in the occurrences of the weather regimes between 2081–2100 and 1961–2000 (number of days during summer) in the CMIP3 models. The changes are given by the length of the shaded bar that correspond to each regimes. The number on the graph stands for the model. Ensemble refers to the ensemble mean of the CMIP3 models.](image)

We have shown that our initial paradigm holds for mid-latitude summer circulation patterns through the 21st century. These results stress the importance of a better understanding of the dynamics of the natural modes of the summer large-scale atmospheric circulation and of their sensitivity to external forcing. We have also shown that they are responsible of the major fraction of the summer precipitation change over Western Europe.

### 3.3 The role of ocean dynamics on NAO variability under global warming conditions

The role of ocean circulation on the NAO-related variability in the Euro-Atlantic sector is investigated. In a previous study (Bellucci et al., 2008) the NAO-ocean circulation interplay was inspected in a simulation of the 20th Century climate (20C3M) performed with the SINTEX-G coupled GCM.

In the present work, we extend this analysis to idealized global warming experiments, under double and quadruple CO2 concentration conditions (2XCO2 and 4XCO2) with respect to preindustrial CO2 levels. The analysis is performed over approximately century long integrations, for stable radiative forcing conditions (i.e., after the transient periods).

It is shown that, while the 20C3M simulation displays a damped oscillatory mode with a typical 5 years time-scale, involving significant lead-lag covariance between the NAO and oceanic heat and mass transport (Bellucci et al.,2008), the evidence of a
similar low frequency mode vanishes when the system is forced towards a warmer climate condition.

In figure 3.3.1a-c, composite maps of winter (JFM) SST anomalies, based on high minus low phases of the SST leading principal component (PC1), at different time-lags, are shown for experiments 20C3M, 2XCO2 and 4XCO2. The SST-tripole pattern (correlated with a positive NAO phase), shows a 5 years re-emergence timescale in 20C3M, with the propagation of SST anomalies appearing to play a relevant role in driving a tripole sign-reversal, after 2-3 years (Fig. 3.3a,d). Experiment 2XCO2 SST-tripole displays a ~2 years decorrelation timescale, with a non-significant re-emergence after 5 years (Fig 3.3b,d), whereas a rapid 1-yr decorrelation is exhibited by the 4XCO2 tripole pattern (Fig. 3.3c,d).

Thus, the predictive skill associated with the re-emergence of the tripole pattern in the 20th century disappears under global warming conditions. Lead-lag covariance between the NAOI and winter SST is also consistently reduced in 2X- and 4XCO2 (not shown), compared with 20C3M.

If the upper ocean-NAO interactions in double and quadruple-CO2 scenarios fit within the white noise paradigm postulated by Frankignoul and Hasselmann (1977), with the ocean acting as an essentially passive player (with respect to atmosphere) in the mid-latitude coupled system, a most active role for the ocean dynamics is envisaged in the 2XCO2 (but not in the 4XCO2) experiment when the coherency between NAO variability and ocean energy transport is examined. In figure 3.3.2, the lagged correlation between ocean meridional heat transport (across several zonal sections in the North Atlantic domain) and the NAOI index is shown, for 20C3M and the two global warming simulations. Interestingly, under double-CO2 conditions, some (marginally) significant coherency between the dominant atmospheric variability mode and the ocean heat transport is recovered, with a sub-decadal timescale similar to the 20th century case, for both lead and lag time intervals. The 2XCO2 correlation map reveals a north-south propagation pattern, suggesting the existence of a coupled ocean-atmosphere interaction which involves coordinated NAO/large-scale ocean circulation changes, which is absent in the 4XCO2 case.
Figure 3.3.1: Composite maps of winter (JFM) SST anomalies (in °C), based on years where the normalized SST PC1 (tripole index) is high and low, with respect to a threshold of one standard deviation, for a) 20C3M, b) 2XCO2 and c) 4XCO2. At lag 0, the composite is simply the difference between the average SST computed over high minus low tripole index. At lag +1 yr (and similarly for larger lags) the composite is computed by shifting the high and low tripole index years by 1 year. d) Lagged autocorrelation of the SST first principal component. Circles denote 95% statistically significant values.

Figure 3.3.2: Lead-lag correlation index between NAOI and meridional ocean heat transport in the North Atlantic domain. White contours encircle the regions which are statistically significant at the 95% level.
3.4 Atlantic MOC internal variability and its response to global warming

The thermohaline circulation (THC) is a global three-dimensional belt of ocean currents that transports large amounts of heat and freshwater (Manabe and Stouffer 1999). In the North Atlantic, it is manifested in a meridional overturning circulation (MOC), which, through its northward transport of warm tropical waters by the Gulf Stream and North Atlantic Current, effectively contributes to the warming of Northern Europe.

Model results indicate that global warming may lead to a weakening or even total collapse of the MOC, which may have serious consequences not only for northern Europe but also for the entire global climate system (e.g., Manabe et al. 1991; Stocker and Schmittner 1997). Whether significant changes in the MOC are already detectable is a controversial debate in the current literature (Bryden et al. 2005). Therefore, more reliable projections of present day ocean circulation and future climate change are essential.

Using an ensemble of 16 different coupled climate models performed for the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC), the evolution of the MOC during the twentieth and twenty-first centuries is analyzed by combining model simulations for the IPCC scenarios Twentieth-Century Climate in Coupled Models (20C3M) and Special Report on Emission Scenarios, A1B (SRSA1B). Earlier findings are confirmed that even for the same forcing scenario the model response is spread over a large range. However, no model predicts abrupt changes or a total collapse of the MOC (Figure 3.4.1).

![Figure 3.4.1: MOC rates at 30°N of all models used for the analysis. There is considerable spread between individual models, even in the year 1900, which is still close to initial conditions. Half of the models match observation-based mass transport estimates for present day (14–18 Sv), which is indicated by the black bar. Dashed lines indicate models where the average of two or more ensemble runs were used.](image)

To reduce the uncertainty of the projections, different weighting procedures are applied to obtain “best estimates” of the future MOC evolution, considering the skill of each model to represent present day hydrographic fields of temperature, salinity, and pycnocline depth as well as observation-based mass transport estimates. Using different methods of weighting the various models together, all produce estimates that the MOC will weaken by 25%–30% from present day values by the year 2100; however, absolute values of the MOC and the degree of reduction differ among the weighting methods (Schneider et al. 2007).
Figure 3.4.2: Weighted best estimates of the MOC projections according to the different skill/weight calculations performed. Even though there is some spread between the different best estimates, all of them show a reduction of the future MOC of about 25%, which is remarkably constant between the different approaches. Furthermore, except for the global Taylor model assessment, all weighted projections for present day fit within the range of observed mass transport estimates [i.e., by this means each of the individual methods (except Taylor global) seems to be justified].

The assessment of model performance, improved the reliability of future climate projections, as uncertainties from individual model simulations are effectively reduced (Figure 3.4.2). Temperature T, salinity S, pycnocline depth (PD), and observed mass transport estimates are suitable parameters to evaluate model performance. Skill calculations should be based on a combination of correlations (Taylor skill) and rms errors of T, S, and PD as well as observation-based mass transport estimates, as no method has turned out to be superior to the others and so different independent approaches are included. According to our findings, a decrease of the MOC over the last 40 yr as described by Bryden et al. (2005) is not seen in any of the models used here, but the future MOC is predicted to decline by about 25% during the twenty-first century. However, additional effects of freshwater input from ice melt (e.g., Greenland) are not considered. The predicted future reduction of the MOC leads to a reduced northward heat transport in the Atlantic by about 7%. This is not sufficient to avoid greenhouse warming over Europe, which will warm during the twenty-first century by about 3 K on average based on these model results.

Superimposed on long-term trend, there is a strong multidecadal variability of the Atlantic MOC. This variability during the twentieth century was investigated by making use of a characteristic relationship between MOC and SST found in global climate models. The simulated Atlantic SST response to multidecadal changes in the MOC is the interhemispheric dipole pattern, with opposite changes in the North and South Atlantic. An index of the MOC during the twentieth century was derived by computing the observed SST difference between the North and South Atlantic. This index exhibits pronounced multidecadal variability during the twentieth century. The evidence was found that these multidecadal MOC variations can be understood as the lagged response to the multidecadal variations in the NAO and the associated variations in Labrador Sea convection (Figure 3.4.3). It was estimated from the analysis of ocean model simulations that the observed density change over the period of 1970–2000 in the region of the Denmark Straight translates into an MOC
change of about 1 Sv, which is well within the range of the natural multidecadal variability that we estimated to amount to about \( \pm 1.5\text{–}3 \text{ Sv.} \)

**Figure 3.4.3:** Time series of the winter [December–March (DJFM)] NAO index (shaded curve), a measure of the strength of the westerlies and heat fluxes over the North Atlantic and the Atlantic dipole SST anomaly index (°C, black curve), and a measure of the strength of the MOC. The NAO index is smoothed with an 11-yr running mean; the dipole index is unsmoothed (thin line) and smoothed with a 11-yr running mean filter (thick line). Multidecadal changes of the MOC as indicated by the dipole index lag those of the NAO by about a decade, supporting the notion that a significant fraction of the low-frequency variability of the MOC is driven by that of the NAO. (top) Annual data of LSW thickness (m), a measure of convection in the Labrador Sea, at ocean weather ship Bravo, defined between isopycnals \( \sigma_{1.5} = 34.72\text{–}34.62 \), following Curry et al. (1998).

While the combined evidence from surface observations, hydrographic data, and model simulations suggests that the variations in the MOC over the last decades can primarily be regarded as a response to the NAO variability, a continuing freshening (or warming) trend in the Nordic Seas must be considered as a key additional factor for the future evolution of the MOC in view of anthropogenic climate change. A recent assessment of global warming simulations by Schweckendiek and Willebrand (2005) suggests that twenty-first century projections of the MOC are mainly tied to the evolution of hydrographic conditions in the Nordic Seas. Most global climate model projections for the twenty-first century suggest a gradual anthropogenic weakening of the MOC of up to 40% (Gregory et al. 2005; Schneider et al. 2007). As described above, the level of internal multidecadal variability has been estimated to about 1.5–3 Sv. Thus, such a weakening will not exceed the range of multidecadal variability within the next several decades (Latif et al. 2006).
References:


Merryfield, W. J. 2006: Changes to ENSO under CO2 doubling in the IPCC AR4 coupled climate models. J. Climate 19, 4009-4027


Yeh, S. W. and Kirtman, B. P. 2007: ENSO Amplitude Changes due to Climate


Zelle, H., van Oldenborgh, G. and Dijkstra, H., 2005: El Niño and greenhouse warming: results from ensemble simulations with the NCAR CCSM. J. Climate, 18, 4669-4683