EMULATE deliverable D12: results of model experiments to determine if the observed relationships in D7 and D11 can be reproduced or can be better resolved using the longer timescales of the coupled model experiments, and an initial study of mechanisms and potential predictability.

Hadley Centre (partner 2), UBERN (partner 6) , SU (partner 7), CEA (partner 4)

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1. Introduction

The role of the oceans and anthropogenic climate change in determining extratropical atmospheric variability is far from clear. Reproducibility of Atlantic and European climate anomalies in the 3rd Hadley Centre model (Pope et al. 2000) forced with observed sea-surface temperature (SST) and climate forcings was therefore a focus of the EMULATE project.

Here we describe results from numerical models and compare them with observed signals. We use two ensembles of simulations run specifically for the EMULATE project and made available to the EMULATE project partners through www.HadC20C.org. We have used a "natural" forcing ensemble which has boundary forcing from observed sea surface conditions, volcanic aerosol and solar variability, as well as an "all forcing" ensemble. This ensemble, in addition to the natural boundary forcings also contains changes in well mixed greenhouse gases (CO_2, CH_4, N_2O, $CFCl_3$, CF_2Cl_2), tropospheric ozone and stratospheric ozone changes since 1975, surface albedo and vegetation changes and anthropogenic sulphate aerosol changes. We present a summary of our main results on the prominent modes of climate variability in these simulations. We use a variety of analysis methods ranging from simple composite analysis, to the application of the new clustering technique on atmospheric data that was developed earlier in EMULATE and applied to observational data (deliverable D7, Philipp et al., 2006).

2. Winter North Atlantic Oscillation (NAO) in EMULATE experiments, additional perturbation experiments and long coupled simulations

Both the natural and all-forcings model ensembles capture the winter NAO as the first mode of winter interannual variability and winter clusters of observational data show positive and negative NAO-like clusters. We therefore projected the modelled daily weather patterns onto the observed clusters (Philipp et al. 2006) by designating each model day according to its nearest cluster centroid. In the January to February period there are clusters which correspond to the negative and positive phases of the NAO. Fig.1 shows that while these anomalies are not exactly symmetric opposites, they do show similar characteristics with opposite anomaly centres over the Azores and Iceland.

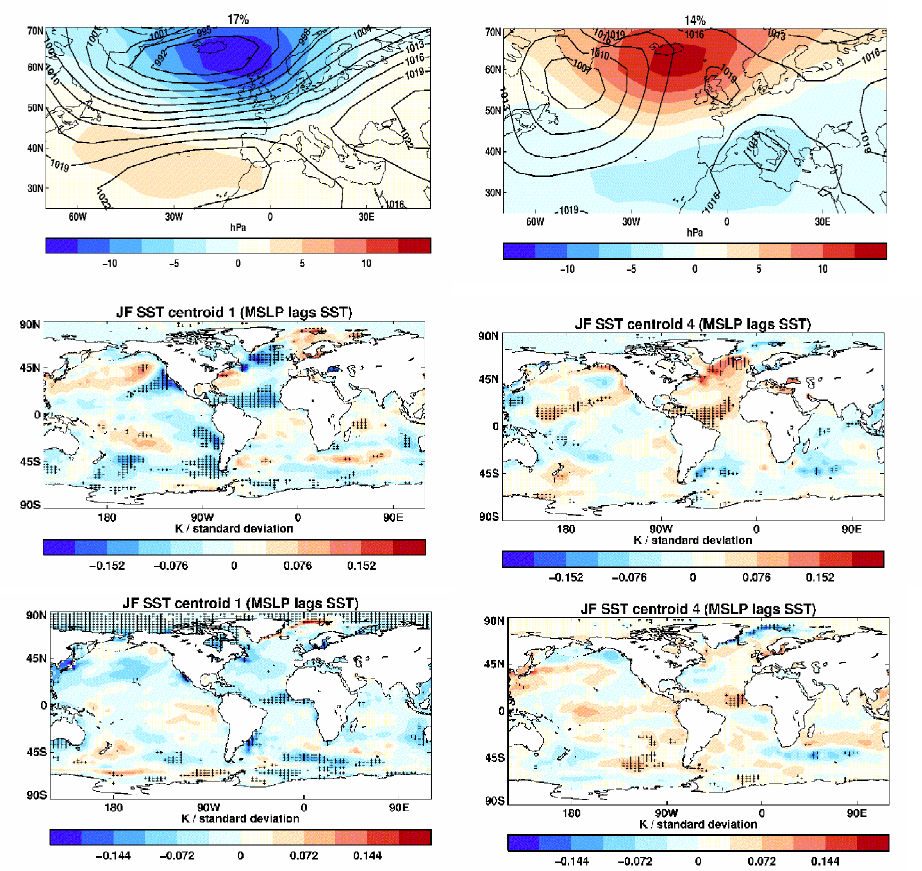

Figure 1: Cluster centroids corresponding to the winter NAO positive phase (left) and negative phase (right) in hPa. Upper panels show sea-level pressure anomalies (coloured) and total sea level pressure (contours), middle panels show observed composite SST anomalies preceding atmospheric anomalies by 1 month and lower panels show similar SST anomalies composited using the modelled cluster frequencies for 1871-2002. Crosses indicate statistical significance at the 90% level.

The modelled interannual variability of the NAO in both the natural and all-forcing ensembles of simulations, and the modelled NAO index of Azores minus Iceland MSLP both show reasonable amplitude when compared to the EMULATE sea level pressure dataset. However, individual year to year variations of the NAO such as the strongly negative NAO in 1962/63 are not reproduced in the model. There is also a striking absence of a strong link with SST on multidecadal timescales. Fig.1 shows that a tripole like SST pattern occurs prior to both the positive and negative phases of the NAO in observations but that this link is only weakly represented in the model.

This weak link between SST and the NAO in the EMULATE ensembles is also seen in the multidecadal trend of the NAO over the latter part of the 20th century. Fig.2 shows that despite including a comprehensive set of radiative forcings, the observed increase in the NAO can not be reproduced in our standard GCM simulations.

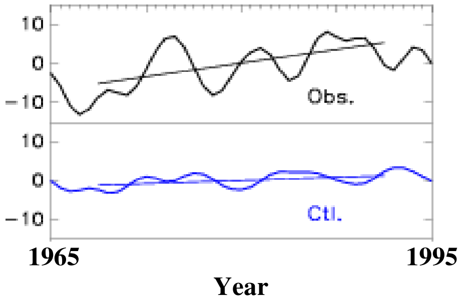

Figure 2: Observed and modelled NAO index (hPa) in a set of GCM simulations with all (anthropogenic and natural) forcings and observed sea-surface temperature and sea-ice.

We also examined the NAO in a long simulation with the coupled ocean-atmosphere version of our model (Gordon et al., 2000) and found a deficit in the low frequency variability of the NAO in these simulations. A typical section of this simulation is shown in Fig.3. It therefore seems that there is a general deficit of multidecadal NAO variability in the model as found by some other authors for other models (Osborn 2004, Kuzmina 2005).

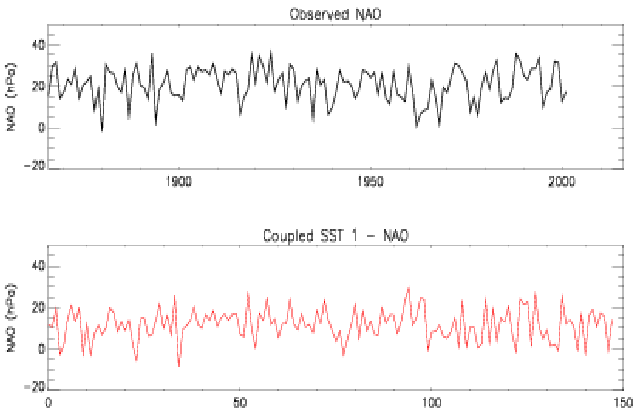

Figure 3: NAO indices (MSLP difference between Azores and Iceland) in observations (upper) and a coupled ocean atmosphere simulation (lower).

Some authors have argued that the lack of reproducibility of NAO variations, and hence European winter climate anomalies, simply reflects the large amount of internal variance in atmospheric extratropical circulation and the predominance of atmosphere-to-ocean rather than ocean-to-atmosphere forcing (Bretherton and Battisti, 2000).

However, there is also evidence that upper level winds may have a strong link to the surface NAO (e.g. Boville 1984, Norton 2003) and we note that the observed trend in the zonal wind at 50hPa and 60N was also not reproduced in our standard ensembles of simulations for EMULATE. The observed trend was close to 7m/s over the period 1965-1995 and the modelled trend was close to 1m/s. It could be that the poor stratospheric resolution of our model is responsible for this discrepancy as our GCM had just 3 model levels above the 75hPa level. In a separate piece of work, we therefore conducted a pair of experiments with imposed upper level circulation changes. By applying a drag on the zonal circulation in the stratosphere of our model which decreased in magnitude with time we were able to reproduce a trend of 8.5m/s in the winds at 50hPa and 60N between 1965 and 1995, in reasonable agreement with observations.

A surprising and important finding is that these perturbed simulations also successfully reproduced the 1965-1995 changes in the winter surface NAO and European surface climate (Scaife et al., 2005). It therefore seems that predictability of winter European conditions could in principle be limited by poor simulation of stratospheric conditions. Additional effects on climate extremes and links with EMULATE work package 4 are documented below.

5. Summer NAO in EMULATE experiments

The observed SNAO for July and August has been compared with an ensemble mean representation of the SNAO from the six integrations of HadAM3 with all forcings run from 1871-2002 (Fig.4).

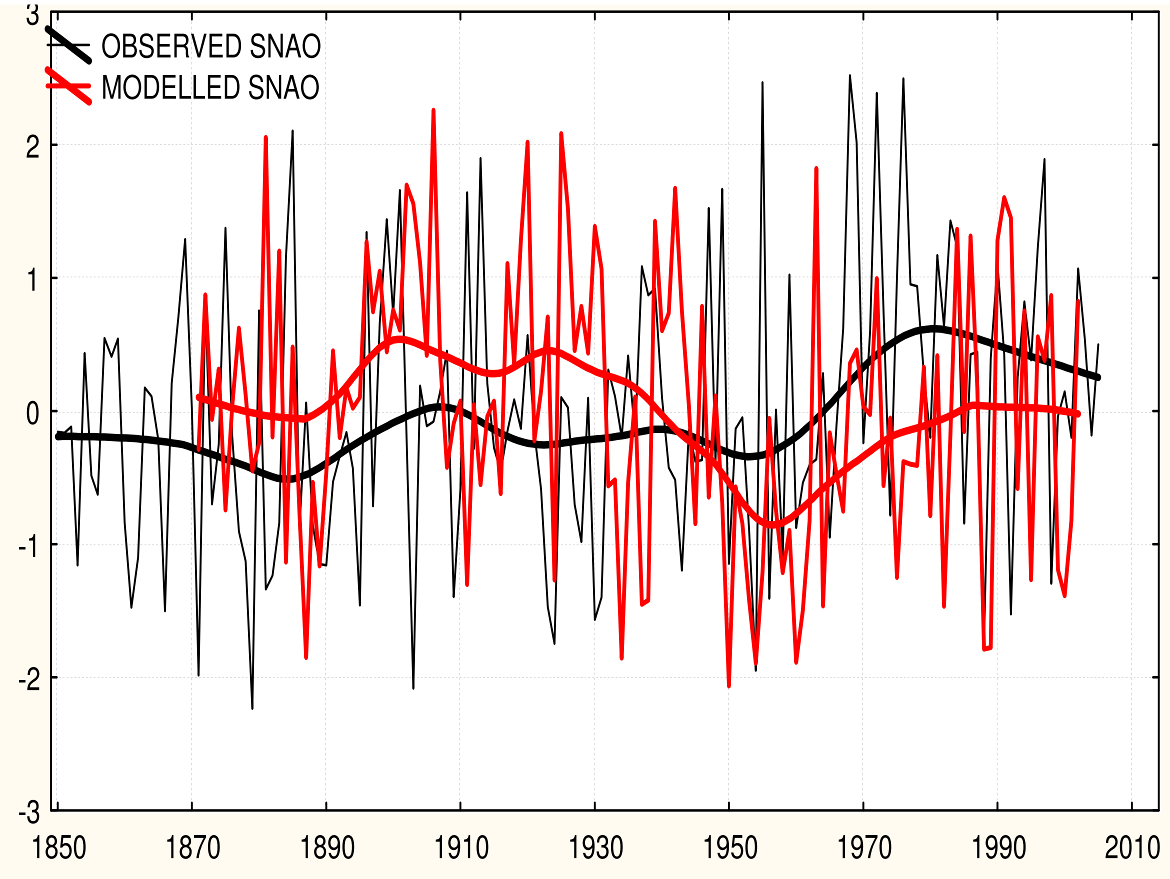

Figure 4: Modelled and observed Summer North Atlantic Oscillation. Observed data are plotted for 1850-2005, modelled data are for 1871-2002. Each annual value is the July-August mean of the daily SNAO EOF coefficient in all individual model runs. The model time series has been normalised to have unit variance over 1871-2002 as has the

observed time series. Model and observed annually- resolved time series, together with a smoothed series are based on a locally averaged regression method. This effectively provides a near bidecadal filter and highlights the low frequency variations.

Concentrating first on the low frequency variability, broadly similar behaviour can be seen between model and observations with a minimum in the SNAO around 1950 in both the observations and the model. This peaks in the observations around 1980 and declines slowly afterwards; a similar peak in the model occurs about a decade later. The late 1960s to early 1970s is a period of marked increase. Fig. 5 shows a regression of the modelled SNAO with sea-surface temperature. It strongly resembles the Atlantic Multidecadal Oscillation SST pattern seen in the observations (see deliverable D7). The relationship has also been calculated for the six runs with natural forcings (not shown). The relationship is qualitatively fairly similar but is stronger and more significant for the all forcings runs.

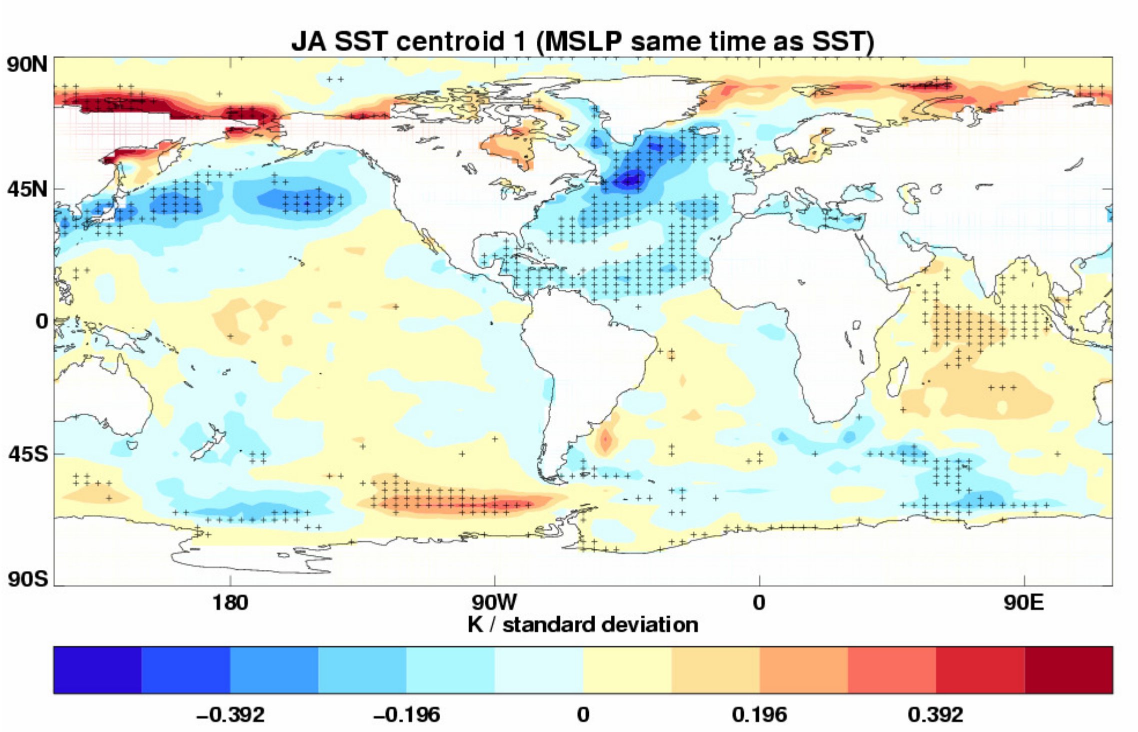

Figure 5: Regression of decadally filtered SNAO against decadally filtered SST where the climate change signal has been removed from the SST. Crosses show points significant locally at the 10% level using a Monte Carlo method.

The extent to which the modelled and observed SNAO series are similar on different time scales is shown by a cross spectral analysis. Here we show the coherence squared (equivalent to correlation squared) between the modelled and observed SNAO as a function of period (Fig. 6). Values greater than about 0.45 are significant at the 5% level. This shows that although the decadal variations of Fig. 4 are overall poorly correlated, there is a significant correlation near the 4 year time scale. This could be related to ENSO as the modelled SNAO and the observed SST in Nino 3.4 have a similar coherency (Fig. 6).

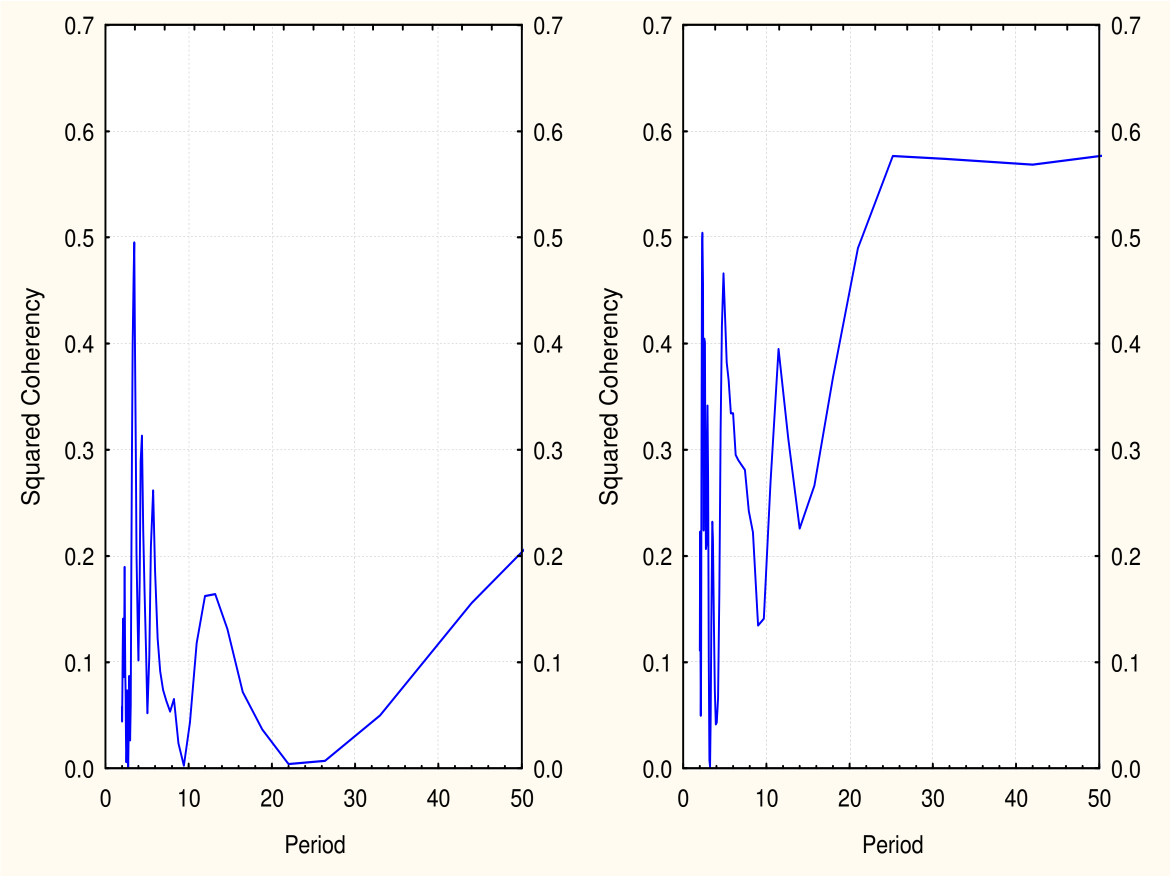

Figure 6: (left) Squared coherency between modelled SNAO and Nino 3.4 SST 1876-2002, (right) squared coherency between observed and modelled SNAO.

We have already suggested there could be a relationship between the observed SNAO and ENSO in D11, so the model could be picking up some of the ENSO signal that influences the SNAO. If this is the case, interannual predictability of the SNAO may exist, as well as the interdecadal signal seen in both the observations and the model that relates to the Atlantic Multidecadal Oscillation. However, the details of this vary between the model and the observations. These preliminary results suggest that some novel predictability may be present for a pattern of atmospheric circulation, the SNAO, which has a strong influence on rainfall and thus drought in high summer over North West Europe.

6. The effect of ENSO on Europe, observations vs GCM experiments and cyclone tracking.

In deliverable report D7 we presented results on the possible effects of the El Nino Southern Oscillation influence on European winter climate. Previous authors have attributed the lack of a robust remote response to ENSO to "non-stationarity", in other words to sensitivity of the remote response to the epoch chosen (Sutton and Hodson 2003, Greatbatch et al. 2004). The implication is that changes in the climatological background conditions on which the remote response to ENSO develops are different in different epochs and that this affects the remote response. However, this has not been clearly demonstrated. An alternative hypothesis is that the response may depend on the amplitude of the ENSO event itself, as different epochs contain different proportions of weak and strong ENSO events. By carefully compositing the weak and strong El Nino events together (Toniazzo and Scaife, 2006) we were able to produce a robust pattern of anomalies over the Atlantic European region as shown in Fig.7.

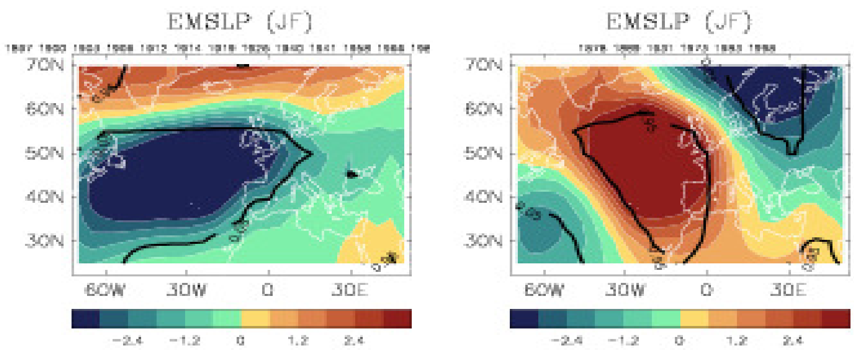

Figure 7: Composite mean sea level pressure anomalies in weak (left) and strong (right) ENSO events from the EMULATE MSLP dataset, January-February means are plotted.

Some aspects of the strong ENSO signal are reproduced in the EMULATE ensemble. Fig.8 shows similar composites of strong and weak events from the model simulations. Although weaker than the observed signals, the strong ENSO case reproduces the high pressure anomalies over the Atlantic and low pressure anomalies over Northern Europe. There is also a tendency towards and extension of high pressure anomaly across the Atlantic in the strong ENSO case but the weak ENSO anomaly shown in Fig.7 is not well reproduced. We went on to investigate the response to ENSO in a set of experiments parallel to the EMULATE ensemble but using SST anomalies corresponding to composite means of weak and strong ENSO events. We noted that non-linearity could arise through the amplitude of the SST anomalies or their pattern which is also different between weak and strong events. Four ensembles of simulations were therefore run, with perturbations to Pacific SST corresponding to each of the combinations of weak and strong ENSO pattern and weak and strong ENSO amplitude.

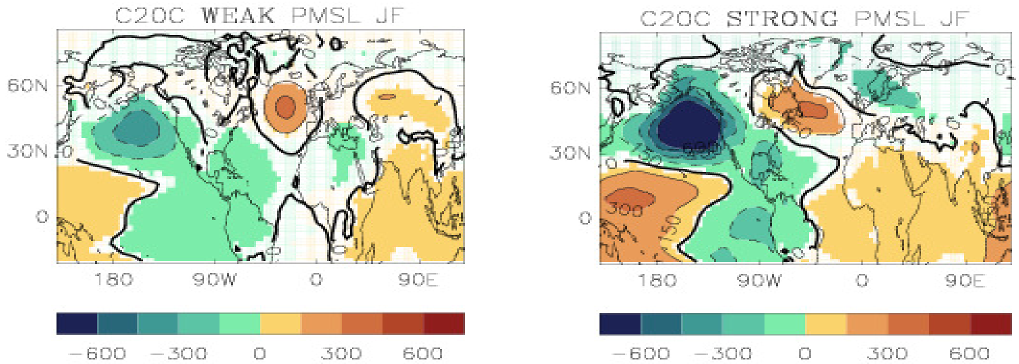

Figure 8: Composites of MSLP anomalies (Pa) in weak (left) and strong (right) ENSO events from the EMULATE ensembles of simulations.

We found that the response is quite linear in our model and that neither the pattern nor the amplitude of Pacific SST anomalies was enough to reproduce the difference between weak and strong ENSO signals seen in the observations in Fig.7. Rather, it seems that differences between tropical Atlantic SST in the strong and weak ENSO cases is most likely to be responsible for the European signal and that the non-linearity acts via an atmospheric bridge from the tropical Pacific to the tropical Atlantic region. Further work is needed to understand this aspect of the ENSO teleconnection.

We also planned to study the influence of ENSO on the Atlantic storm track using tracking algorithms on the individual low pressure centres in the EMULATE observational and modelling datasets. The algorithm of Murray and Simmonds (1991) has been developed for use with data with coarse spatial and temporal resolution (5x5deg, daily). It has been run on the EMULATE sea level pressure data and all of the long model simulations done for EMULATE. Extremes of the pressure field are identified iteratively starting from maxima (minima) of the Laplacian of the pressure field. This allows for open lows (maxima in relative vorticity without a related pressure minimum; i.e. lows within troughs) to be tracked as well. Tracking is carried out as follows: For all tracks separately, the position of the lows is estimated using a combination of climatological cyclone movement and cyclone movement since the last analysis time (if available). This estimate is compared with computed cyclone positions and the nearest cyclone is chosen as the next member along the track (taking into account numerous constraints such as change in central pressure, distance between estimated and observed position, possible assignments of the cyclone to other tracks, etc.). Finally, gridded statistics such as system density, velocity of lows, central pressure, central tendency, etc., are computed based on the output of the tracking procedure. The parameter presented here is system density. System density is measured in percentage of systems/area (a system density of 100 means either one cyclone present within the specified area (25 deg.lat.squared for EMSLP) at all times or 100 systems present within area at one time.

We found large discrepancies between the modelled and observational climatologies (Fig.6) which prevent a useful comparison of the two. Errors in the EMSLP datset in the region south of Greenland lead to large track densities while the model simulated cyclones show low track density mainly due to inadequate model resolution to produce deep cyclones. Both of these factors make any further comparison of modelled and observed storm tracks difficult.

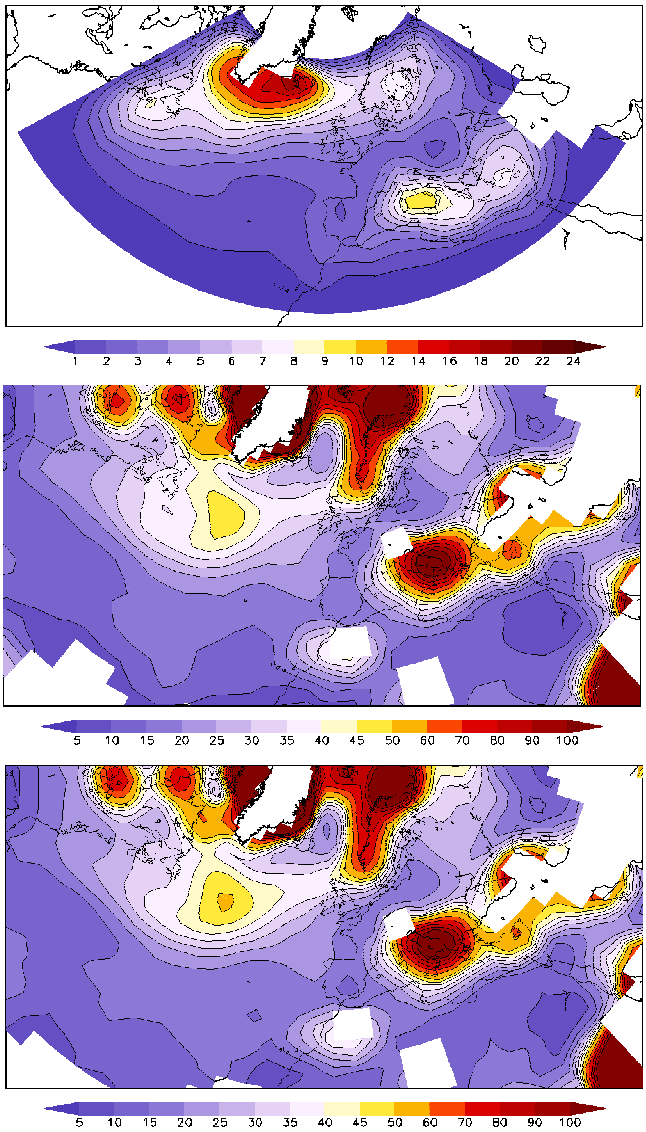

Figure 9: Climatological winter (DJF) cyclone track densities for 1881-2000. Climatology of system density (in units of percentage of systems/25 deg.lat.squared) for EMULATE sea level pressure data (upper), all ensemble members with natural forcings (middle) and all ensemble members with natural and anthropogenic forcings (lower).

7. Atlantic Multidecadal Oscillation and ultra-long timescale predictability.

In earlier EMULATE work (deliverables D7 and D11), it was noted that some of the cluster frequencies appeared to be related to an interhemispheric contrast pattern in

Atlantic sea surface temperature. This pattern has also been identified as a long lived and low frequency mode of natural climate variability (Vellinga and Wu, 2004, Knight et al., 2005) and is termed the Atlantic Multidecadal Oscillation (AMO). This appeared to be most robust in summer but it also suggested that the analysis ought to be extended to longer simulations and then to be re-examined in other seasons of the year. To do this we have examined a multicentury simulation with the coupled ocean-atmosphere version of our model (Gordon et al., 2000). We use a method outlined in a separate study (Knight et al., 2005) to isolate the AMO and we have now examined modelled sea level pressure and precipitation signals over the EMULATE region (Fig.10). Clear, statistically significant impacts of the AMO are found over the European region, suggesting a decrease in sea level pressure over Europe and the Atlantic at all seasons of the year. Associated with this decrease in pressure is a tendency for increased precipitation over Europe and the Sahel region of Africa. The response over this latter region can be understood in terms of changes in the position of the inter-tropical convergence zone (Folland et al., 1986) or both its position and intensity (Rowell et al. 1992) but the mechanism for the summer change over Europe is less clear and requires further work.

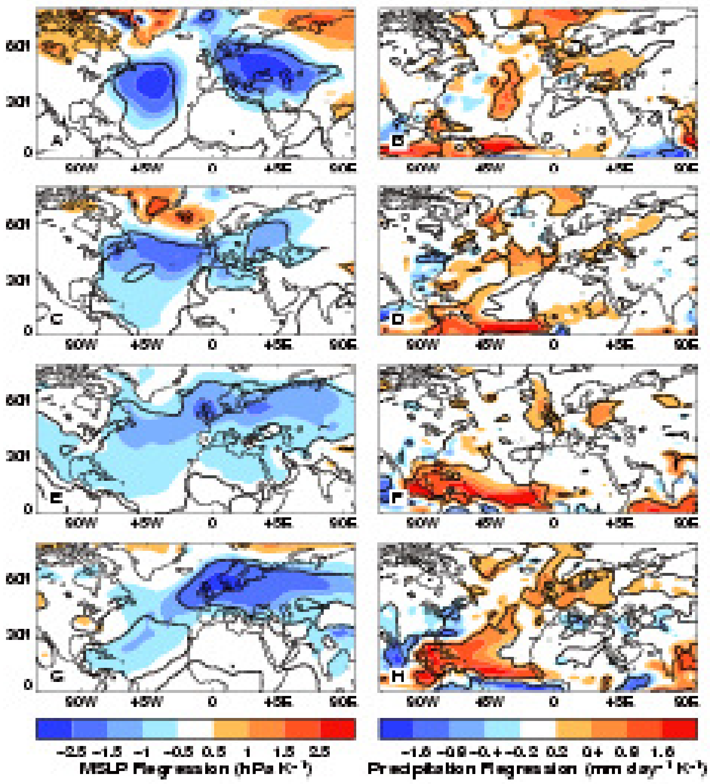

Figure 10: Simulated seasonal impact of the AMO on extratropical circulation and precipitation. Regressions of the AMO onto MSLP (left) and precipitation (right) are shown for DJF, MAM, JJA and SON (upper to lower). After Knight et al. (2006).

There are consequences for decadal predictability of European climate in this work. It has already been established that the AMO shows some predictability out to at least a few decades; because although the oscillation is irregular, it does show coherency on timescales of up to 50 years (Knight et al., 2005). Given the significant associated signals in European seasonal mean climate, the AMO therefore represents a new (albeit weak) potential source of decadal forecasting predictability for Europe.

8. Extremes and links with EMULATE WP4

The modes of variability documented in this report are not only responsible for variability in the mean climate but also cause variations in climate extremes. In addition to the WP4 analyses of trends in climate extremes we have also examined the response of European climate extremes to the main mode of extratropical winter variability i.e. the NAO.

As the EMULATE ensemble (like other models) was able to reproduce only a small fraction of the observed increase in the NAO between the 1960s and 1990s, this ensemble was used as a "control" for comparison with the simulations described in section 1 in which upper level circulation changes gave rise to a large increase in the NAO. The result of this increase in the NAO on heavy rainfall events over Europe is shown in Fig.11.

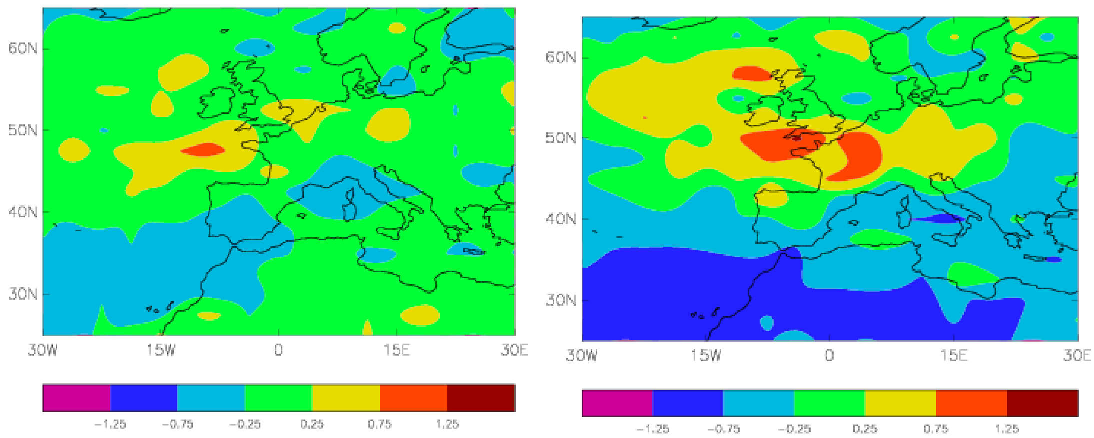

Figure 11: Changes in the frequency of heavy rainfall events between 1965 and 1995. Fractional changes in the frequency of 90th percentile rainfall are shown between 1965 and 1995 in the control EMULATE simulations (left) and simulations with increasing NAO (right)

Changes in rainfall events in both the control and increasing NAO simulations have a dipolar structure. This corresponds well with an increase in *mean* rainfall in northern Europe and the decrease to the south associated with the NAO (not shown). The changes are also large, with 75% changes in frequency in areas such as north west Europe. The area mean values north and south of 42N in Fig.8 are also approximately 5 times larger in the simulation with increasing NAO than they are in the simulation with only greenhouse gases and other climate forcings included. In the light of these modelling results, further evidence for the impact of the NAO on extremes is also being sought using the observational station datasets produced in WP4. Fig.12 shows one such example where the relationship between the NAO and the occurrence of European heavy rainfall events is verified over the full century timescale. Further examples are being shown in a report by Mohammad et al. (2006). This provides a clear demonstration that changes in dynamical modes of variability and not just radiative forcings are crucial in explaining observed changes in climate extremes on regional scales (Scaife et al. 2006).

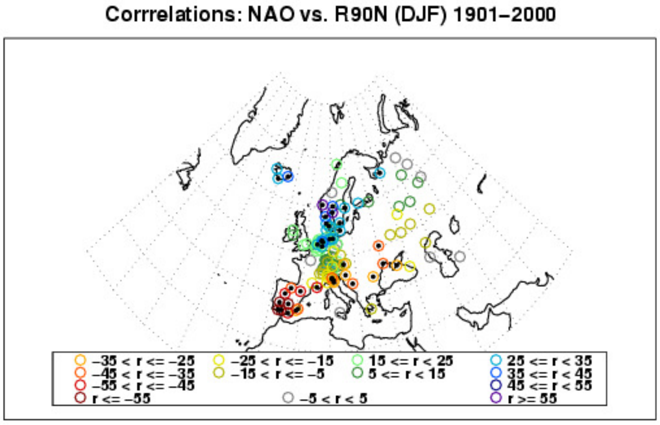

Figure 12: Correlations between the frequency of 90th percentile daily rainfall and the North Atlantic Oscillation over the 20th century using EMULATE station data.

9. Conclusions and recommendations for future work

This report describes a range of aspects of climate variability in the EMULATE model simulations. We have carried out an extensive analysis of our model simulations using daily cluster analysis and a comprehensive paper is being written on the comparison of modelled and observed clusters (Fereday et al. 2006).

It turns out that for our model, like many others, even specifying observed global sea-surface temperatures and other climate forcings is not a sufficient condition to simulate more than a fraction of the observed increase in the winter NAO (c.f. Cohen et al. 2005). However, in parallel experiments to those carried out for EMULATE, we have shown that the increase in the NAO can be reproduced in models if upper level circulation changes are included. This suggests that future models used to simulate historical European conditions ought to include an improved representation of the stratosphere and this question will be answered under continued work under the FP6 EU DYNAMITE project. Current work is also demonstrating significant potential forecast skill from models on decadal timescales (Smith et al., 2006) and use of extended models should also be tested for possible benefits to long range forecasting skill on seasonal to decadal forecast times.

We also tried to characterise the remote effects of ENSO on the Euro-Atlantic region. Using the long EMULATE MSLP dataset provides a clear advantage for this type of study over previously available and much shorter records. Further modelling work is required to attempt to reproduce the observed winter signals, perhaps focussing on tropical Atlantic SST anomalies. Similarly, higher model resolution may be needed to accurately examine cyclone statistics and compare these to observations. Modelling experiments with localised SST anomalies would also be useful to test the proposed relationship between ENSO and the NAO in summer.

Finally, there are consequences of this work for interpretation of past climate change signals over Europe. When annual trends are broken down into seasonal trends, great care must be taken to properly account for modes of climate variability. For example, the Atlantic Multidecadal Oscillation can easily project onto regional trends in a variety of meteorological variables yet we know this is a natural variation of climate. Similarly, although we can not be sure that the observed change in the winter NAO is not anthropogenic, recent observations show a downturn in the NAO which is consistent with natural variability. If the NAO variations are natural in origin then great care must be taken in interpreting observed trends in both mean and extreme climate events. If, on the other hand, the NAO variations are anthropogenic in origin, then improved models are needed which better represent stratospheric processes; in this case we could be underestimating the rate of winter climate change expected from anthropogenic forcing. Either way, understanding changes in the primary modes of climate variability is a key question for the next few years.

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