TECHNICAL MEMORANDUM

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Decadal Variability: processes, predictability and prediction

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Abstract

The climate system exhibits variability on a variety of timescales. Over much of the ocean the spectra of sea surface temperature and salinity are red, broadly consistent with the ocean acting to integrate atmospheric white noise forcing. However, there are substantial areas where this is not an appropriate approximation and other processes are operative. Various modes of climate variability such as the Atlantic multidecadal oscillation (AMO) and the Pacific Decadal Oscillation (PDO) have been suggested. The thermohaline circulation in the Atlantic is identified as a key component of the AMO through the Atlantic Meridional Overturning Circulation (AMOC) though the details of how they are related are still unclear. Various processes are identified and their predictability assessed. Predictability horizons have been estimated using numerical models and preliminary predictions made and skill assessed. In general there is some hope for predictability in the Atlantic sector on time scales up to about ten years arising from knowledge of the ocean initial state. In the last year or two a beginning has been made to develop decadal prediction systems. The approach is somewhat similar to that developed over the last decade for seasonal forecasting. In addition to initialising the ocean, greenhouse and other radiatively active gas concentrations are used. There is a large scatter in the forecasts, partly arising from errors in the models and uncertainties in the ocean initial conditions.

Although decadal prediction is in its infancy, predictability studies indicate some encouraging potential. Large-scale low frequency variability such as temperature over Europe or rainfall over the Sahel offer the potential for huge benefits to society. Delivering accurate reliable prediction is a major challenge for the next decade.

1 Introduction

Over much of the globe the ocean sea surface temperature changes broadly in line with the white noise integrator idea of Hasselmann (1976). This is embodied in the simple equation below

$$\partial T/\partial t = -\lambda T + \zeta \tag{1}$$

where ζ represents forcing of the ocean by the atmospheric heat flux and T represents sea surface temperature (SST).

If the atmospheric noise is white, T in eqn 1, has a spectral form:

$$G(\omega) = \sigma^2/(\lambda^2 + \omega^2) \tag{2}$$

where σ is the standard deviation of the atmospheric noise, and so the ocean will have a red spectrum. The λ term represents a feedback term of the atmosphere on the ocean. Equation 1 can hold for sea surface salinity (SSS) too, driven by net fresh water forcing (precipitation minus evaporation), but λ would be smaller because the atmospheric feedback on salinity is considerably weaker than that on temperature. Since the details of the high frequency forcing are not predictable beyond a few days, eqn (1) indicates only a modest opportunity for forecasting low frequency variability.

Fig 1a shows spectra from ocean weather station Papa in the north Pacific. Both temperature and salinity have red spectra, with the spectrum for salinity being considerably redder than that for T, exactly as one might expect from the local linear stochastic forcing argument. However, Fig 1b from weather station India in the north Atlantic shows a rather different result. While the spectra for T and S are both red and indicate some turning at around 5-10 years, they do not show salinity to have a redder spectrum than T. In fact the spectra are remarkably similar. Analysis of coherence by Hall and Manabe (1997) show that the two fields are highly coherent at periods longer than about 10 years, indicating something other than local stochastic forcing is at work.

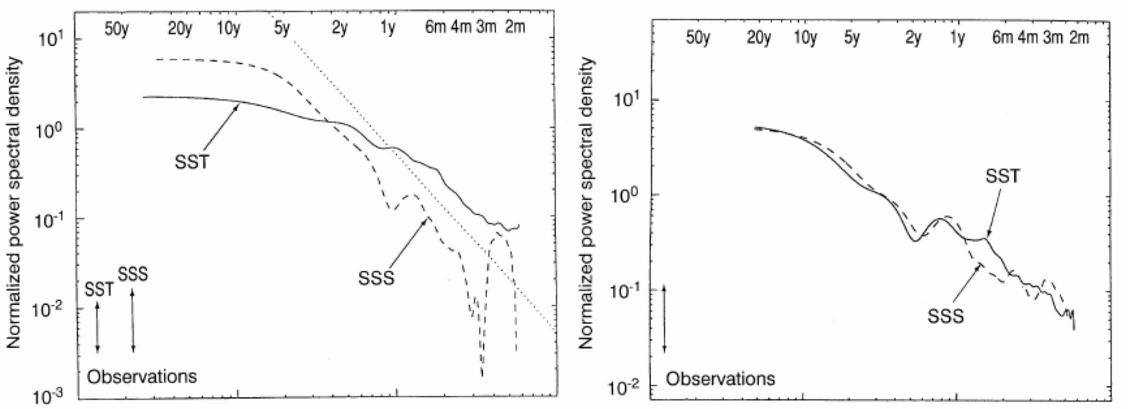

Figure 1: Spectra of sea surface temperature and sea surface salinity at ocean weather stations Papa (North Pacific) and India (North Atlantic). The spectra for Papa are consistent with local stochastic forcing by the atmosphere with the atmospheric spectrum being white. A different story applies in the North Atlantic, where the T and S spectra are very similar. From Hall and Manabe (1997).

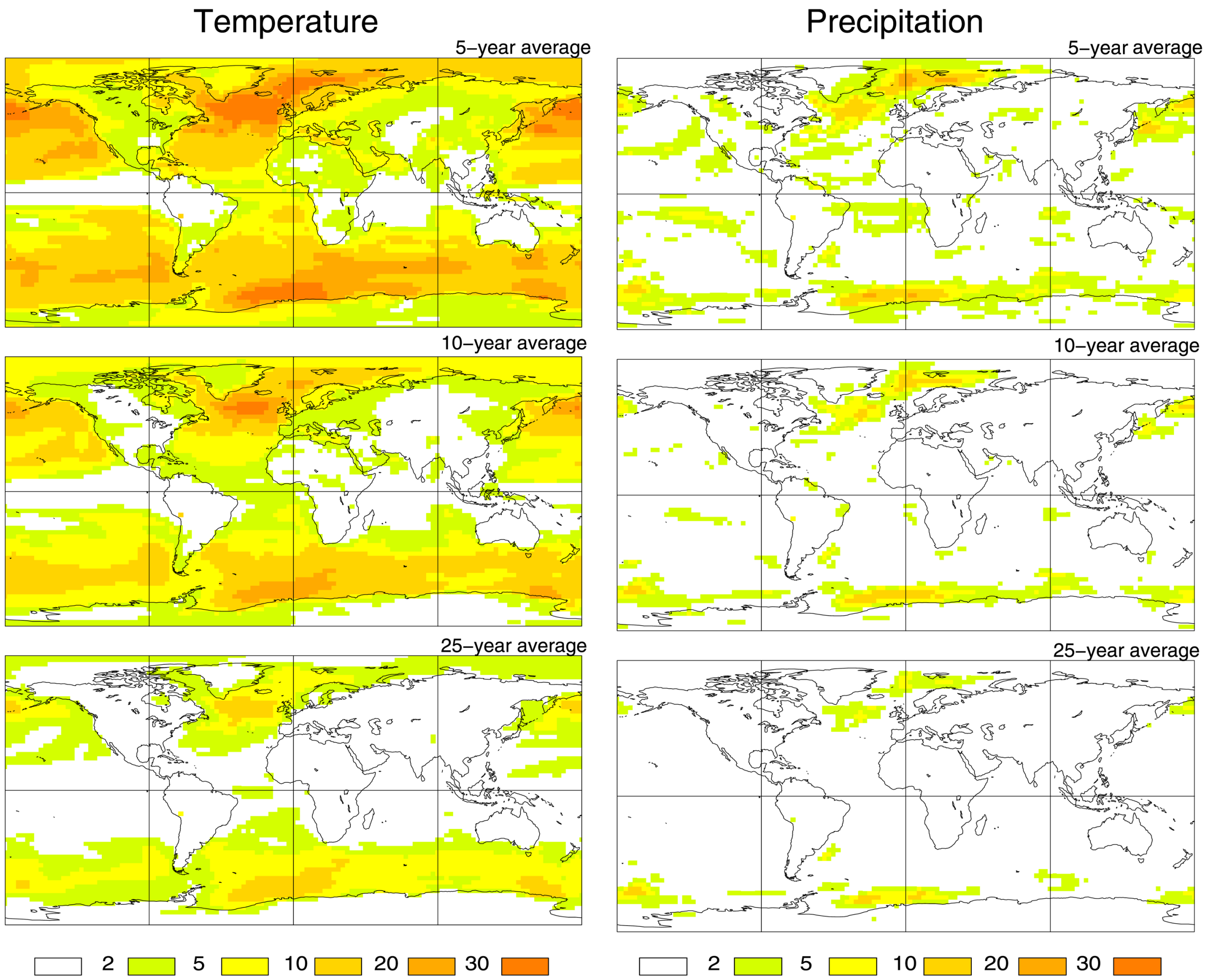

Figure 2: Estimated potential predictability variance fraction (ppvf) expressed as a percentage for 5, 10 and 25 year averages for temperature (left) and precipitation (right). Potential predictability of precipitation is much less than that for temperature though the patterns are similar. Over land potential predictability is small for both temperature and precipitation. From Boer and Lambert (2008).

There are insufficient data to allow spectra of SST and SSS to be evaluated other than at a few locations (mainly weather ships) but Hall and Manabe (1997) go on to consider the global picture based on coupled model results and show that while local stochastic forcing is likely to apply quite widely as a 'cause' of low frequency variability, there are regions where it is not applicable: in particular, the North Atlantic, the North-East Pacfic, the tropical Pacific and the southern oceans polewards of 50°S. Frankignoul et al. (1997) look in greater detail at the decadal response of the North Atlantic to stochastic forcing, but concentrate on fields other than SST, such as surface current response to wind stress forcing. They conclude that their extended stochastic forcing model, including eastward propagating long Rossby waves can account for much of the decadal variability in pressure and currents.

Saravanan and McWilliams (1998) have also refined the simple stochastic model of Hasselmann (1976) to include an equation for atmospheric temperature similar to eqn (1), to include a meridional advection term in the ocean, as well as an interaction term in both atmosphere and ocean to represent feedback between the two media. Analytic solutions are posible for specified advection velocity. An interesting feature of this model is that preferred timescales of variability arise in the ocean response, even if there is no intrinsic oscillatory mechanism in either fluid. The time-scale depends on the ocean advection velocity and the length-scale of the atmospheric forcing. The estimated timescale for the North Atlantic is around 10 years.

The existence of regions where there is some indication of processes beyond integrating white noise, opens the possibility of forecasting low frequency variability. These areas of potential predictable skill are usually estimated on the basis of model studies. Some processes which can lead to forecast skill have been identified based on observational studies, but in general the temporal and spatial coverage on which observational studies could be based is limited. This is especially true of the subsurface ocean which plays a key role in setting the mechanism, location, and timescale of low frequency variability. Consequently many results discussed below are based on model studies. In reality, data are needed to check the validity of the models but in practice the models are inadequately verified. Different models may have different mechanisms and different preferred frequencies for low frequency variability but the data are usually too sparse to allow discrimination of one mechanism in favour of another. Nonetheless, as will be shown later in Fig 4 there are indications of potentially predictable signals in the North Atlantic, South Atlantic (including most of the Circumpolar current), North Pacific, and tropical Pacific, broadly consistent with the regions indicated by Hall and Manabe (1997). The degree to which conditions over land are predictable is more debatable but even here there may be some limited, though still possibly useful, predictability.

An alternative estimate of predictability can be obtained by analysing the variability from a single model run or from observations though the record is too short to be definitive. The record is split into a low frequency part and internally generated noise part. The idea is that the former may be associated with processes which may be predictable. Fig 2, from Boer and Lambert (2008), shows a recent estimate of both temperature and rainfall predictability for different averaging periods. What is plotted is ppvf (potential predictability variance fraction), which is the ratio of the low frequency variance to the total variance. The results for T are broadly consistent with classical predictability studies such as that of Hall and Manabe (1997) and other studies discussed later, in that the highest predictablity is in the mid to high latitudes especially the north Atlantic. Predictability over land is small because the high frequency (unpredictable) variability is much larger over land than over the ocean. There is a hint of predictability in the vicinity of the Himalayas, probably related to snow cover but there is little predictability in the Equatorial Pacific on 5-year or longer averages. Although there is considerable predictability associated with ENSO, this is on annual or shorter timescales. The right hand columns show estimates of predictability of precipitation. This is broadly similar in pattern to T but much weaker. As noted by Boer and Lambert (2008) the largest values of ppvf are associated with regions where the surface layers of the ocean make contact with the deeper ocean beneath. The results shown are based on nearly 9000 years of model integration using 21 different models. The data set is therefore large and so the estimates are likely to

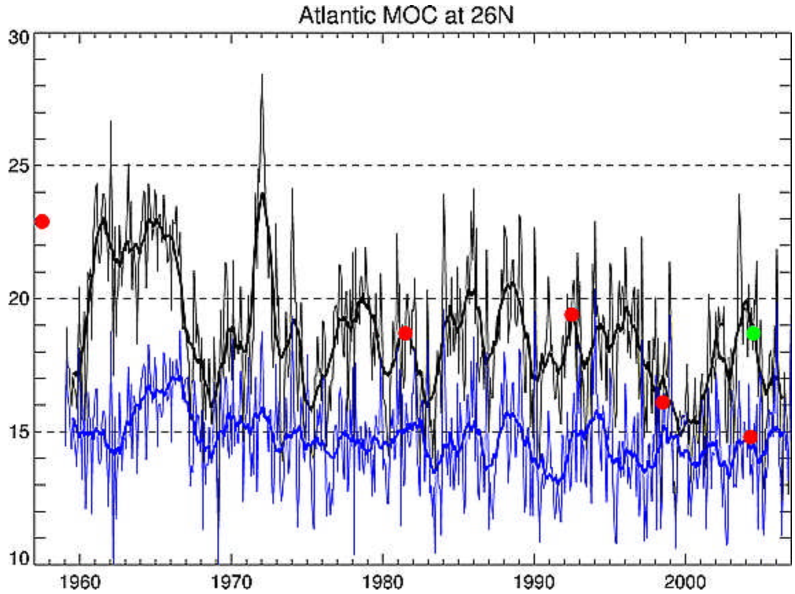

Figure 3: Atlantic meridional overturning circulation over the last 50 years as a function of time. Also plotted are values from trans-ocean transects at 26°N. The error bars are likely to be large on the observed values and the inaccuracy of the model estimates is also significant. Observations, indicated by red dots are from Bryden et al. (2005) and the green dot is from Cunningham et al. (2007). Black indicates results from the ECMWF S3 analysis system (ORA-S3) in which all available data are assimilated. Blue indicates results are from a forced ocean run in which no ocean data other than SST are used. From Balmaseda et al. (2007).

be statistically significant. The patterns are broadly similar but are not the same as those obtained by classical predictability studies whereby one sees how quickly ensemble members diverge in a perfect model scenario. This will be discussed in greater detail later.

In Section 2 we describe some basin scale processes pertinent to the North Atlantic and North Pacific, including interactions between them and relating processes to the meridional circulation in the ocean as a potential mechanism for low frequency variability. The discussion of the North Atlantic Oscillation (NAO) is limited as this review is already long and a moderately recent review of the NAO exists (Marshall et al., 2001). In Section 3 we discuss predictability resulting from both natural causes as well as resulting from changes in the concentration of radiatively active gases. In recent years, decadal prediction systems have begun to be developed. In Section 4, three of the four extant systems are reviewed. Decadal forecasting is an initial value problem requiring accurate initial conditions for the ocean. Analysing the ocean is difficult and many uncertainties remain, as discussed in this section. In Section 5, preliminary results from the fourth decadal prediction system are discussed within the context of the major EU project ENSEMBLES.

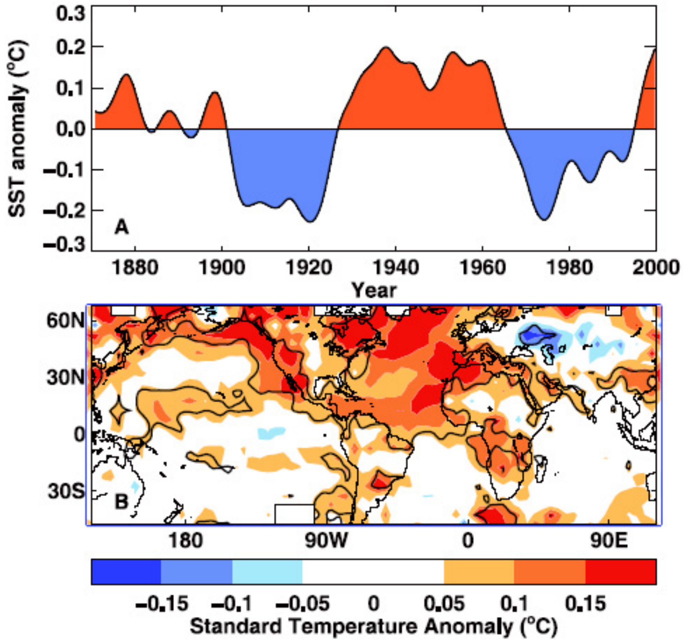

Figure 4: Upper) Plot of the low frequency variability in the North Atlantic SST represented by the AMO index. Lower) Surface temperature anomaly associated with one positive standard deviation of the AMO. Based on observations, from Knight et al. (2005).

2 Some basin-scale processes

2.1 The North Atlantic

There have been times in the past when temperatures in the Atlantic have been much colder than now. Temperature over Greenland in particular has changed rapidly according to paleo records, by many degrees within just a few years. Implicated in these changes are the Overturning Circulation (AMOC) and the Atlantic Multidecadal Oscillation (AMO). The AMOC, even today is difficult to measure and certainly this was even more so in the past. The AMO is identified through surface temperature. It is plausible that the AMO is closely linked to, possibly caused by, changes in the AMOC, a stronger AMOC giving rise to greater northward heat transport in the ocean and warmer SSTs in the North Atlantic. Model studies of the ocean response to increasing greenhouse gases indicate a weakening of the AMOC. This weakening may be gradual but the possibility of some threshold being reached and a rapid irreversible slowdown of the AMOC can not be ruled out (Dijkstra et al., 2006). The AMOC is almost impossible to measure directly but estimates can be obtained indirectly from ocean analyses. One example, shown in Fig 3 from Balmaseda et al. (2007) indicates a slight weakening of the AMOC in recent times. The estimate is from the ORA-S3 ocean reanalysis (Balmaseda et al., 2008). 'Observed' values based on ocean transects at 26°N are also shown. The red dots are from Bryden et al. (2005) which suggested a possible reduction in the strength of the AMOC; the green dot is from Cunningham et al. (2007), considered to be more representative than the Bryden et al value for that year. This value is higher than the Bryden et al. (2005) value and suggests a weaker downward trend in the AMOC strength. There is no strong evidence that the AMOC has increased in the last decades, but see Eden and Willebrand (2001) and Knight (2009). The error bars both for the observed values and for model estimates are large as will be discussed later.

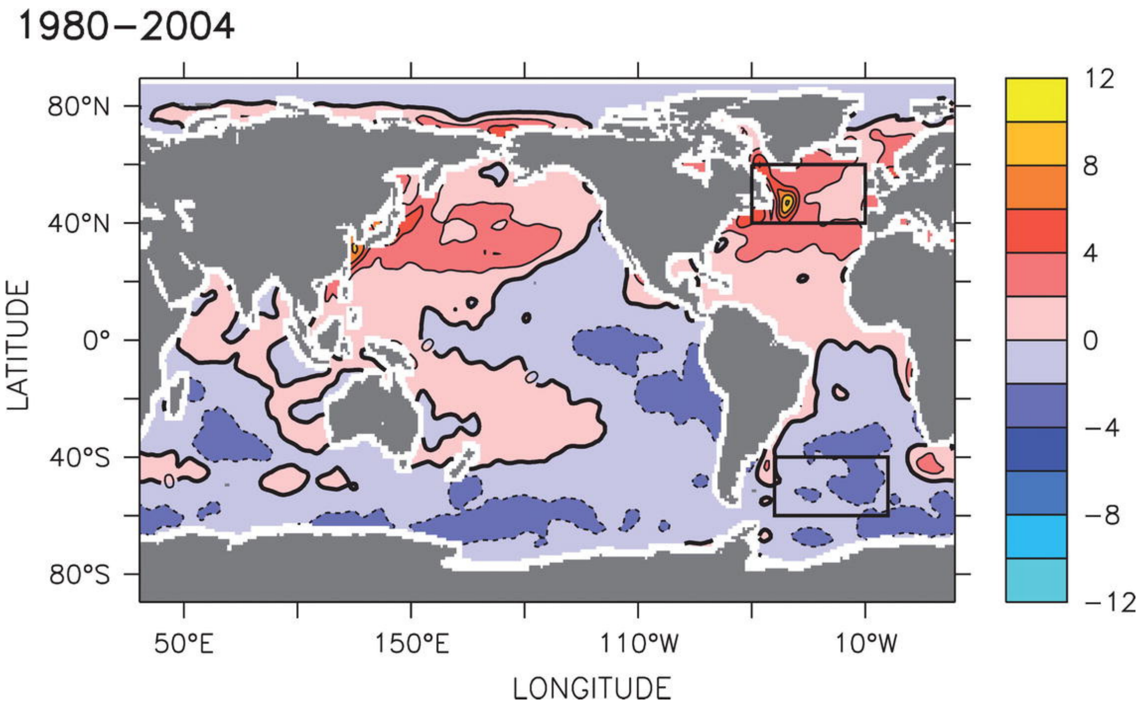

Figure 5: Plot of the trend in SST in K/century for the 25 years 1980-2004, after the global mean trend has been removed. The dipole structure of the AMO is more evident here than in Fig 4. Note also the signal in the North Pacific. From Latif et al. (2006).

Fig 4 from Knight et al. (2005) shows a time series of the amplitude of the AMO over the last 130 years. This is similar to the analysis of Sutton and Hodson (2005) though there is heavier smoothing in the latter. The AMO has been linked with Sahel drought, the frequency of hurricanes, North American and European climate (Knight et al., 2006, Pohlmann et al., 2006). The AMO switched into a warm phase in 1996. In order to draw a link between the AMOC and the AMO, Knight et al. (2005) use a long climate run (1400 years) using constant values of external forcing (Gordon et al., 2000). Analysing a spectrum of SST in the North Atlantic from their model, only one peak stands out and that is at low frequency, at a period around 100 years. Using wavelet analysis they show that the variability varies considerably from one epoch to another. There are periods when the AMO is strong and others when it is weaker but it is always present suggesting that it is a robust oscillation in the model. The period is a little longer than would be estimated from observations of the last hundred years or so but generally the model AMO bears a fair resemblance to the observed AMO. Paleo data also indicate that the AMO is a robust feature of the past climate. See for example Grosfeld et al. (2007). Through model studies Knight et al. (2005) link the AMO variability to changes in the Atlantic ThermoHaline Circulation (ATHC) and find a strong link between poleward heat transport at 30°N and the ATHC, in the form of the AMOC. However, the observed SSTs have warmed considerably since the mid 70's (Fig 4) at a time when one would expect, based on anthropogenic forcing arguments, the AMOC to be weakening, not strengthening. On the other hand Eden and Jung (2001) argue, based on the low frequency variability of the North Atlantic Oscillation that the AMOC should be strengthening. This is currently an unresolved issue suggesting that the link between AMOC and AMO or at least North Atlantic SSTs is not well understood. An interesting and possibly relevant point here comes from Balmaseda et al. (2007) who showed that although the AMOC had slowed somewhat in their analysis, the reduction in heat transport was considerably smaller because the thermal contrast between upper poleward waters and lower equatorward flow had increased, so offsetting the reduction in flow rate.

In a later paper, Knight et al. (2006) corroborated many of the observational links between the AMO and climate such as the multidecadal variability in North-East Brazil rainfall, Atlantic hurricanes, North American and European climate. Based on observations, the warm phase of the AMO (1930s-1950s) is linked to decreased North-East Brazil rainfall and USA rainfall and enhanced Sahel and Atlantic hurricane formation. By using a 1400 year integration of the coupled model HadCM3 it is possible to put any relationships on a more secure statistical footing and if the model faithfully represents the AMO and its connections, these results can be used to increase confidence in the observational links. They find that the link between the AMO and Brazilian rainfall is well simulated and linked to shifts in the spring position of the ITCZ. Likewise reduction in Sahel rainfall occurs in the negative phase of the AMO due to a southward shift of the ITCZ in summer. The amplitude of the reduction in Sahel rainfall in the model is much less than observed and probably is indicative of model error. The model also shows a link betwen the vertical shear in the hurricane development region and the AMO as well as a link with low frequency variability in the tropical Pacific. Increased cyclonicity and rainfall over parts of the Atlantic are simulated. These are present year round but most noticeable in summer. The patterns are consistent with Sutton and Hodson (2005) in that there is more rainfall over the Atlantic in the warm AMO phase. The signal over north west Europe is broadly consistent with Sutton and Hodson (2005) but that in US rainfall is not. Nor is it consistent with observations (decreased US river flows during the warm AMO phase in the 1930s-1950s).

Alternative ways of defining or identifying the AMO have been used. Fig 5 shows the trend in SST in degrees per century for the period 1980 to 2004. The global mean trend was removed. This procedure, or one based on eofs in which the first EOF is the global warming trend, may tend to emphasise the Atlantic dipole structure of the AMO, rather than emphasise the North Atlantic as in Knight et al. (2005). Patterns of variability are also dependent on the chosen period. A comprehensive coherence analysis has not been done but it appears that the Atlantic tripole pattern is important on interannual timescales and the interhemispheric dipolar pattern or the northern part of it on decadal timescales, but emphasis switches to the hypermode discussed by Dommenget and Latif (2008) for multi-decadal variability. It is characterised by activity in the Pacific, the northern part

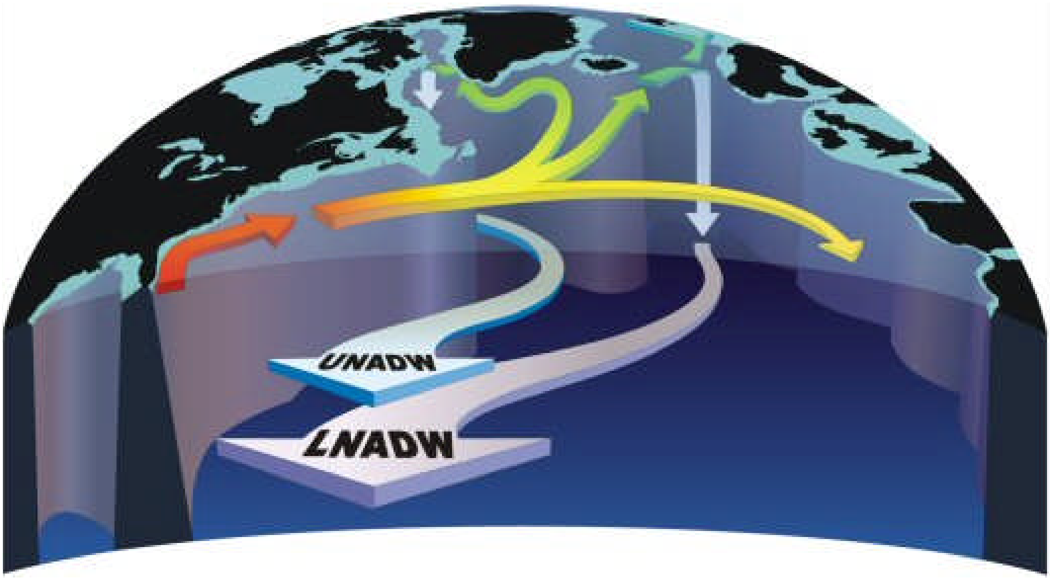

Figure 6: Schematic of the major currents in the North Atlantic. The northward flowing warm Gulf Stream water is cooled in the Labrador and Greenland, Iceland and Norwegian (GIN) seas to form Upper and Lower North Atlantic Deep Water which then flows southward at depth. From National Oceanographic Centre website, http://www.noc.soton.ac.uk/rapid/rapid.php.

being out of phase with the equatorial region, with some connection to the Atlantic and Indian oceans. In long control integrations of IPCC-AR4 climate models, this mode is reproduced and, as in observations, highlights the equatorial and North Pacific. It appears that ocean dynamics are not required for this mode; indeed a model based on the noise forcing idea of Hasselmann (1976) together with vertical diffusion of heat is all that is required to capture most of patterns of variability. Dommenget and Latif (2008) further suggest that the origins of this mode are in the extratropics, atmospheric teleconnections carry the signal to the tropics and from there around the globe. We will not consider the hypermode further in this review, as not much is known of it as yet.

Fig 6 shows a 3-D cartoon of the currents in the North Atlantic. The northward flowing warm Gulf Stream flows along the western boundary in the upper ocean, crosses the Atlantic with branches flowing into the Labrador sea and the Greenland, Iceland, Norwegian (often called the GIN) sea. In the northern seas the 'warm' water is cooled and sinks to form upper and lower North Atlantic deep water respectively. The separation into upper and lower North Atlantic deep water can be clearly seen in observations [1]. The strength and depth of these currents is affected by salinity as well as temperature. [2] Increased fresh water input or reduced cooling would act to slow the AMOC. Some model studies point to the Labrador sea, rather than the GIN sea as being the more important source of decadal variability. See for example Latif et al. (2006). Biastoch et al. (2008), however, suggest that a significant amount of variability in the AMOC in the tropical and subtropical North Atlantic has its origins in the Southern Hemisphere, from the Agulas current.

During the 19th and 20th centuries, there have been marked changes in the *summertime* climate of both north

[1] see for example http://www.noc.soton.ac.uk/OTHERS/woceipo/

[2] The North Atlantic is much saltier than the North Pacific and this is thought to be one reason why the Atlantic has a vigorous overturning circulation and the Pacific does not. The vigorous overturning in turn leads to a warmer north Atlantic through increased heat transport. There is also a greater evaporation from the Atlantic and greater condensation in the cooler Pacific, maintaining the salt contrast.

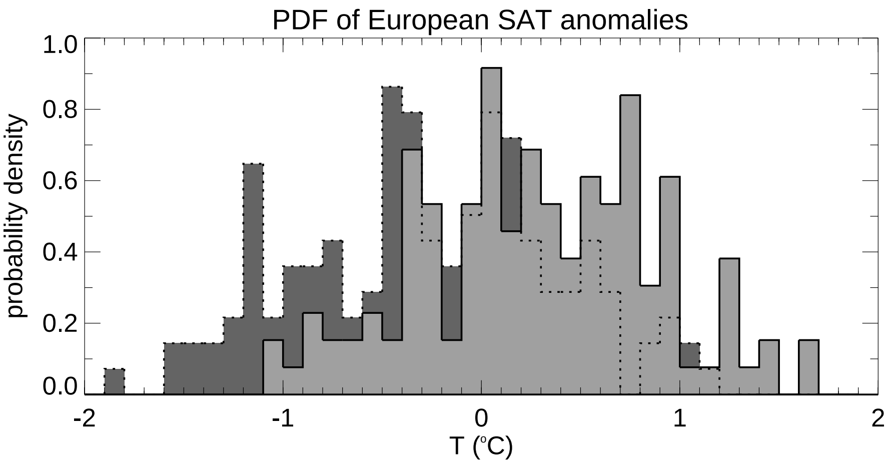

Figure 7: Impact of AMOC on European SAT (surface air temperature averaged over the domain 35°N-75°N, 10°W-40°E). The PDFs of SAT for anomalously strong and weak AMOC states are shown by light and dark shading. From Pohlmann et al.(2006).

America and western Europe, both in terms of temperature and rainfall. It has been suggested that these changes result from changes in the AMOC. Sutton and Hodson (2005) consider changes in the sea level pressure, air temperature and precipitation between a warm phase, 1931-1960, and a cool phase, 1961-1990. See Fig 4. There are two prominent low pressure centres: one over the southern US, associated with reduced precipitation and one to the west of the UK associated with enhanced precipitation. The variations are not dominant but neither are they small, perhaps up to about 20% of the mean value. Through model studies, forcing the HadAM3 atmospheric model with observed SSTs, Sutton and Hodson (2005) provide support that these changes result from changes in North Atlantic SSTs. Through further experiments varying SSTs only in the Atlantic they showed that the influence of the AMO extended into the Pacific and tropics and by further dividing the region into northern and tropical Atlantic, that the climate anomalies in the southern US and Mexico were related to the tropical Atlantic and those in Europe to the extratropical Atlantic. The Atlantic ocean may thus play an important role in decadal variability on a global scale, not just in the Atlantic sector. A link between AMO and ENSO is also possible and will be discussed later. These results are for summer conditions when natural variability in the atmospheric fields considered is weak. In winter conditions when atmospheric variability is larger, it is much harder to show a reproducible relationship. Pohlmann et al. (2006), using a coupled model integrated for 500 years with greenhouse gases set to 1990 levels have investigated the European climate in the model for both strong and weak AMOC conditions. The probability distribution functions (pdfs) of annual mean European temperature are calculated for weak and strong MOC conditions. The difference in the median temperatures is about 0.5K, which is a considerable fraction of the natural variability, as shown in Fig 7.

Djikstra et al. (2005) have considered mechanisms for the AMO, using solutions of a hierarchy of models. In the highest member of the hierarch, (the GFDL-R30 climate model) it is the dominant statistical oscillatory pattern of variability. It is not easy to understand the physics of this variability directly from the GCM. The timescale is set by the basin crossing time of density anomalies, though this is not a priori clear in the GCM. In the lowest member of the hierarchy, it results from destabilisation of the mean thermohaline circulation. By considering other intermediate hierarchy of models, it is possible to trace a connection betweeen the two.

One of the large-scale patterns of atmospheric variability is that of the North Atlantic Oscillation (NAO), which consists of a north-south dipole of pressure anomalies, one centre being located over Iceland/Greenland and the other to the south between 35°-40°N. The NAO index is usually taken as the pressure difference between Reykjavik and Lisbon or the Azores. Both phases of the NAO are associated with large-scale changes in the jet stream over the Atlantic and temperature and precipitation in Europe and North America. The NAO exhibits considerable interannual variability, and the wintertime NAO exhibits considerable decadal variability.

While there is uncertainty about the origin of NAO variability, the picture is somewhat clearer with respect to the ocean response to NAO variability. Dickson et al. (1988) describe changes in the convection and subsurface water mass structure and relate these to changes in the NAO. Curry et al. (1998) linked changes in Labrador sea water to deep sub-tropical gyre water with a timelag of around 6 years. Eden and Willebrand (2001) investigated the mechanisms of interannual to decadal ocean response to realistic surface forcing, highlighting the role of the NAO. They used the forcing (surface stress, heat flux and fresh water flux) from the NCEP reanalysis covering the period 1958 to 1997. The model is able to reproduce much of the observed SST variability, especially in winter with correlations up to 0.8, presumably because the model represents deep winter mixed layers rather better than it does shallow summer layers. The response in the tropical Atlantic is poor. This could be an ocean model problem but they consider that it is probably a reflection of uncertainty in the reanalysis fluxes in this region.

The variability in heat transport at 48°N is compared when the model is forced with full fluxes and when it is forced with that part of the fluxes linked to the NAO. The agreement is very close at low frequency indicating that most of variability in heat transport is linked to the NAO. They also consider some idealised permanent positive and permanent negative NAO experiments. The ocean response to the former is almost the opposite to the latter indicating that although highly nonlinear processes such as convection are involved, the overall ocean response is remarkably linear. They also consider the time variability of the heat transport at 48°N in response to changes in the NAO. The initial response to a positive NAO is a rapid reduction in the heat transport. After 5 years the heat transport has increased to that before the NAO onset, and by 8 years it is significanly higher than it was initially. The mechanisms for these various timescales involving barotropic and baroclinic wave responses is discussed through a series of experiments involving changes to only wind forcing or heat flux forcing. It is unclear whether the coherence between NAO and AMOC holds for longer timescales (Balmaseda et al., 2007).

Eden and Jung (2001) have also considered the influence of the NAO on North Atlantic SSTs and circulation. Using an atmospheric reanalysis, they separate out heat and fresh water fluxes as well as the wind forcing that are linked to the NAO and then force an ocean model of the Atlantic with these forcings from 1865 to 1997. A comparison run using full forcing fields from the NCEP reanalysis is hardly different to one forced with with the reduced forcing leading to the conclusion that the NAO drives most of the low frequency variability. Additionally they find that when the ocean is forced just with the heat flux variability, and climatological values are used for the fresh water and wind forcing, then the response is very similar to using all forcing. From this. they conclude that it is the heat flux which is crucial, especially that in the north west Atlantic in the region of deep water formation. Their Fig 8 which shows the AMOC at 48°N is strikingly similar in the two runs. They also find that the NAO leads the ocean response by several years. What causes low frequency variability in the NAO is not resolved though they argue it is not the AMOC.

Latif et al. (2007) have considered links between the AMOC and the NAO. Based on multi-channel singular spectral analysis they find two modes- one multidecadal (the AMO) and the other quasidecadal. The former explains about 30% of North Atlantic SST variability and the latter about 15%. In their interpretation, multi-decadal ocean variability is a response to multidecadal NAO fluctuations which change the formation of deep water in the Labrador Sea while the quasidecadal mode in the ocean is heavily damped but results from coupled atmosphere ocean interaction. The SST patterns associated with the two modes are different: the AMO is a monopole in the North Atlantic whereas the quasidecadal mode has a tripolar pattern. This interpretation of the two modes is not universally accepted. As indicated by Latif et al. (2007), the AMO and NAO are not clearly linked in all models. The quasi-decadal mode is similar to that discussed by Eden and Greatbatch (2003).

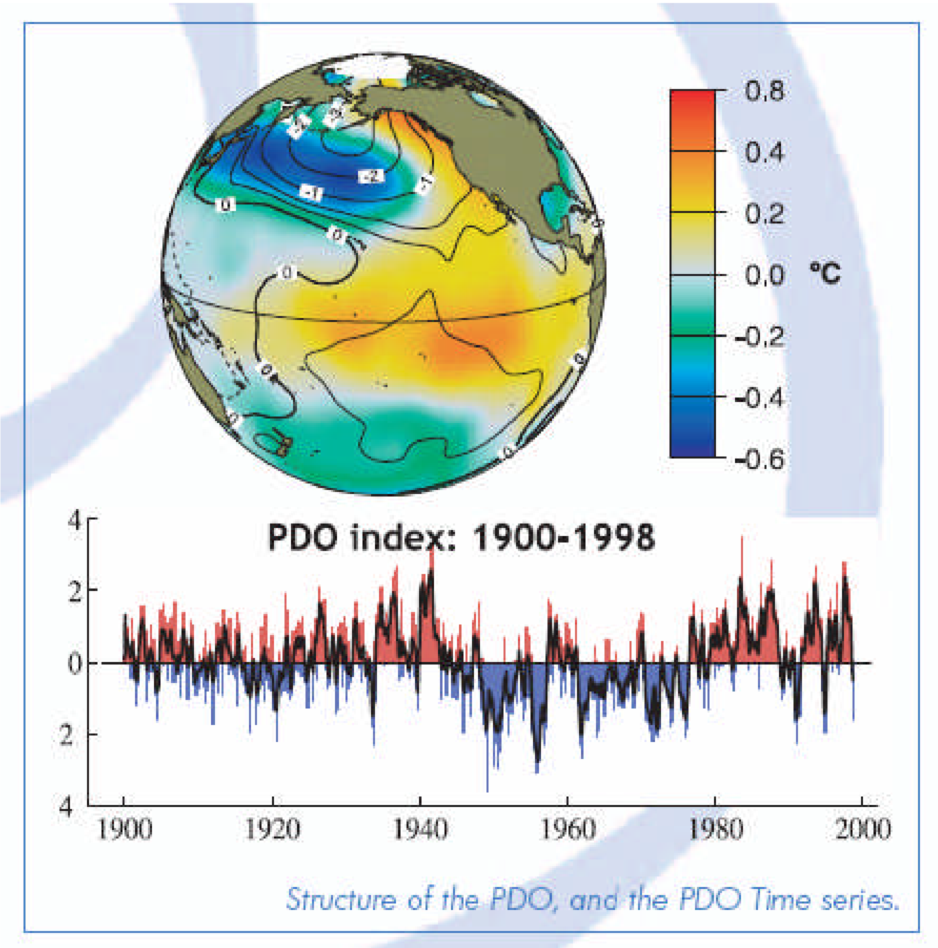

Figure 8: Schematic of the Pacific Decadal Oscillation (PDO) and its time variability. From Hare and Mantua web site: http://jisao.washington.edu/pdo/

2.2 The Pacific Decadal Oscillation

The Pacific Decadal Oscillation (PDO), shown in Fig 8 is a long-lived pattern of Pacific climate variability, with warm or cold periods persisting for 20-to-30 years. It is most visible in the North Pacific/North American sector, with secondary signatures in the tropics in contrast to ENSO which is strongest in the equatorial region with a weaker signal at mid latitudes. Fig 8 shows cool PDO conditions prevailed from 1890-1924 and from 1947-1976, while warm PDO conditions dominated from 1925-1946 and from 1977 to the present (Hare and Mantua, 1996). The link between the PDO and ENSO is a matter of considerable debate, some aspects of which will be discussed below.

Major changes in North-East Pacific marine ecosystems have been correlated with phase changes in the PDO; warm eras are associated with coastal ocean biological productivity in Alaska and inhibited productivity off the west coast of the United States, while cold PDO eras induce the opposite effect (Mantua et al., 1997). The PDO has been linked to decadal variability in the weather patterns over North America and with variations in rainfall over the United States (Latif and Barnett, 1994 and 1996). In a recent paper Woolhiser (2008) found significant influence on precipitation in Nevada but little influence of the PDO on Arizona and New Mexico. Pavia et al. (2008) considered the effects of the PDO and ENSO on the climate of Mexico. Liu and Chan (2008) considered the effect of the PDO on tropical cyclone tracks in the west Pacific where one might expect the PDO to alter the winds, including the steering level winds and vertical wind shear. However, the influence did not seem to be very strong. The PDO has also been linked to the depth of the mixed layer. The variability is related to changes in the local winds. A contrast is made with the North Atlantic where mixed layer variability is not obviously related to wind changes but more to advection changes, consistent with the study of Hall and Manabe (1997). See also White et al. (2008). Causes for the PDO are not currently known, nor is its potential predictability. If the PDO arises from air-sea interactions with multi-year ocean adjustment times, then aspects of the phenomenon may be predictable at lead times of up to several years.

Barnett et al. (1999b), based on five different climate simulations consider much of the mid latitude decadal variability of the North Pacific to be stochastically driven, consistent with Hall and Manabe (1997). They also found a coupled mode of the atmosphere-ocean system with a period of about 20 years which they denoted PDO. They consider several different models, with a range of ENSO behaviour ranging from no ENSO variability, weak ENSO variability, to normal ENSO variability. One model has no ocean-atmosphere interaction at all. In all cases, the mid latitude variability is similar indicating that it is not of equatorial origin. In fact the latter configuration suggests that the origin of the variability lies in the atmosphere, though it may be amplified by interaction with the ocean. They suggest that the mid latitude variability may influence ENSO, rather than low frequency equatorial processes causing the mid latitude variability. This view is not universally accepted, however. See for example Zhang et al. (1997), Zhang et al. (1998) or Deser et al. (2004).

With respect to basin scale variability, Barnett et al. (1999b) find white atmospheric spectra, but red oceanic spectra, consistent with stochastic forcing, as discussed in Section 1. On a regional basis they find the greatest signal around the Kuroshio extension region. The timescale for this is twenty years apparently the same in both the NCAR and Hamburg models (but Latif et al., 2004, analysing a subsequent ECHAM model found the period to be rather longer). This timescale could be set by advection round the gyre and Rossby waves travel times.

Barnett et al. (1999b) consider in more detail the mechanisms of the PDO (i.e. the 20-year variability in the above five coupled models), finding a close relationship between North Pacific SST and equatorial SST with maximum correlation at zero lag. However, they argue that it is only the stochastically-induced low frequency variations in the mid latitudes which projects onto the ENSO region and that the PDO does not. Schneider et al. (1999) consider oceanographic pathways from the mid latitudes to the equatorial region. They find that while upper ocean thermal anomalies appear to propagate from the northern mid latitudes to the equatorial region decadal variability within the tropics is largely independent of the arrival of thermal anomalies from mid latitudes but rather is dominated by tropical wind forcing. This result is inconsistent with the interpretation of Zhang et al. (1998) who interpret the large 1997/98 El Niño to warm water advection from mid-latitudes.

The warming in 1976/77 has received extensive coverage though Zhang (1998) analysing an SST data set spanning 1900-1992 argue that the 1976/77 shift is but one example of the PDO changing sign, others being in the 1940s and 1900s (See Fig 8). It is unclear if the PDO referred to by Rodgers et al. (2004) and concentrated in the tropics is the same as that referred to by most others, in which the variability is located mainly in the North Pacific. To partly escape this, Rodgers et al. (2004) introduce the nomenclature Tropical Pacific Decadal Variability (TPDV) and consider whether TPDV is generated by coupled processes within the tropics or if it involves the extratropics. Even if generated within the tropics it is unclear if it is stochastically excited or the result of nonlinearities in the tropical system. Changes in the equatorial thermocline depth and or stratification can modulate the amplitude and frequency of ENSO. Coupled model studies have shown such links with increased ENSO variability linked to deeper thermocline in the eastern Pacific (Kirtman and Schopf, 1998). Using a 1000 year integration of the ECHO-G model Rodgers et al. (2004) show that it is the asymmetry between warm and cold events that gives rise to low frequency variability. This is supported by observations in the east Pacific but not throughout the equatorial region. The model has a stronger more regular ENSO cycle than in nature and consequently gives a clearer indication of the effects of the asymmetry. In reality there is more noise, perhaps other processes also creating decadal variability and so the effects of the asymmetry are less easy to identify or confirm to be operative in observations.

Further analysis of the interconnection between the North Pacific and the equator is given in Solomon et al. (2003). It is still unclear how North Pacific decadal variability is related to equatorial decadal variability. Gu and Philander (1997) postulated one connection pathway in which temperature anomalies in the extra tropics were subducted and advected to the equator. This pathway is not clearly identified in observations and model studies have suggested that little of the heat anomaly reaches the equator. Given that there is no clear concensus on the origins of decadal variability in either the North Pacific or the equatorial region, it is not surprising that there is no clear understanding of how they are connected.

The North Pacific could in principle sustain oscillations if there is a region of positive feedback between the atmosphere and ocean, with a mechanism to set the timescale and damping of the feedback through advection or Rossby wave propagation. Likewise such a delay oscillator can exist in the equatorial region giving rise to ENSO variability and nonlinear processes can give low frequency variability to ENSO. This does not preclude interaction between the two, however. The role of stochastic forcing also seems to be unclear in maintaining the decadal mode in the North Pacific.

Solomon et al. (2003) have explored the link between tropical and extra-tropical variability in a constructed coupled model. The ocean was a 3.5 layer model, incorporating a surface layer, a thermocline layer and an upper intermediate water layer above an inert deep layer. The atmosphere was greatly simplified. It contained two primary wind patterns linked to SST, derived from the observed correlations, and a boundary layer to calculate heat flux transfer between the two media. In addition stochastic wind forcing based on observed variability can be included. There is no atmospheric link between the extra-tropics and the tropics. The only way the extra tropics and the tropics can interact in this model is through an oceanic pathway.

The model has decadal variability in the North Pacific, based on coupled positive feedback with extra-tropical Rossby waves providing the delayed negative feedback. The model also has ENSO-type interannual variability. If only one of the wind patterns is active then these modes exist independently. When both wind patterns are active, both modes are present and interaction takes place. The interannual mode now exhibits decadal variability, the connection from the extra-tropics to the tropics being via the oceanic subtropical cell. Because the model is constructed, it is possible to specify the strength of coupling between the atmosphere and ocean.

By weakening this coupling, it is possible to have a region in which the extratropics does not show oscillations on its own but when the tropics are active the extratropical decadal mode can be excited. With this choice of parameters the atmospheric connection from the tropical region to the extratropics is important as stochastic forcing in the extratropics is not able to maintain the extratropical decadal mode.

Interesting though their results are, the model is too simple to be definitive. The prescription of the wind patterns as a response to SST variability is a particular weakness, also identified by the authors. Nonetheless this study highlights the potential mechanisms for tropical extratropical interaction at decadal timescales.

More recently Latif (2006) has revisited the PDO and analysed the results of a 2000 year run of the ECHAM3-LSG coupled model. Results are consistent with Barnett et al. (1999a) in that there is no need for tropical-extratropical interaction in order to get low frequency variability in the North Pacific. However, the results are inconclusive as the resolution is poor in both atmosphere and ocean and the ENSO variability is likewise poor. It could be that in nature such interacton is operative and in fact studies such as Deser et al. (2004) would argue that the equatorial region is important. The observational record is short, however, so this study too is not definitive. Latif (2006) argues, as did Barnett et al. (1999a), that much of the mid latitude variability is stochastically forced. Contrary to Barnett et al. (1999a), he argues that the preferred time-scale is set by advection as suggested by Saravanan and McWilliams (1998). There is also controversy over how strongly the atmosphere responds to mid-latitude SST anomalies. The ECHAM3-LSG model used by Latif (2006) has only a weak interaction at mid latitudes whereas the model used by Latif and Barnett (1996) has very strong coupling. The 'correct' level of coupling is unknown. Newman (2007) has considered the PDO as well as shorter term processes and concludes that the potential predictability is not much more than a year.

2.3 Interaction between the Atlantic and Pacific

In a recent study Timmermann et al. (2007) considered the influence of a weakening of the AMOC on ENSO. This is based on several coupled models run in a configuration in which the AMOC is seriously weakened through an injection of fresh water in the northern latitudes of the Atlantic. There are at least two methods by which a slowing AMOC can influence the Pacific; either through an atmospheric bridge or through an ocean teleconnection via coastal and equatorial Kelvin waves. The latter involves a Kelvin wave travelling down the western boundary of the Atlantic to the equator, along the equator, down the west side of Africa round Cape Horn and up to the equator in the Indian ocean, along the equator, through the Indonesian Throughflow and into the equatorial Pacific, potentially extending as far as the South American coast. Convoluted as it is, Timmermann et al. (2005), using a coupled model of intermediate complexity, showed that ENSO variability is suppressed by this mechanism. However, the atmospheric bridge from the Atlantic to the Pacific may have a bigger effect, acting to increase ENSO activity. There is no definitive mechanism linking AMOC and ENSO; the mechanisms may be subtle and the models not yet sufficiently realistic to get all the competing processes correct. There is some fragmentary evidence linking the AMO with ENSO, in that the annual cycle in the Niño3 region was strong and ENSO variability was weak during a period when the AMO was positive 1930-1960 and vice versa when it was negative (1970-1990s). The model study of Dong et al. (2006) reproduced the relationship between ENSO and AMO using the HadCM3 model as shown in Fig 9. This relationship (negative AMO-strong ENSO) was also seen in some of the models used in the Timmermann et al. (2007) study.

2.4 Induced collapse of the AMOC

So far, the discussion has concentrated on natural fluctuations in the Atlantic and Pacific oceans. There have been concerns, however, that the Atlantic MOC, responsible for transporting a large fraction of heat northwards could actually collapse. Stommel (1961) in a landmark, but at the time, largely unnoticed paper, showed that

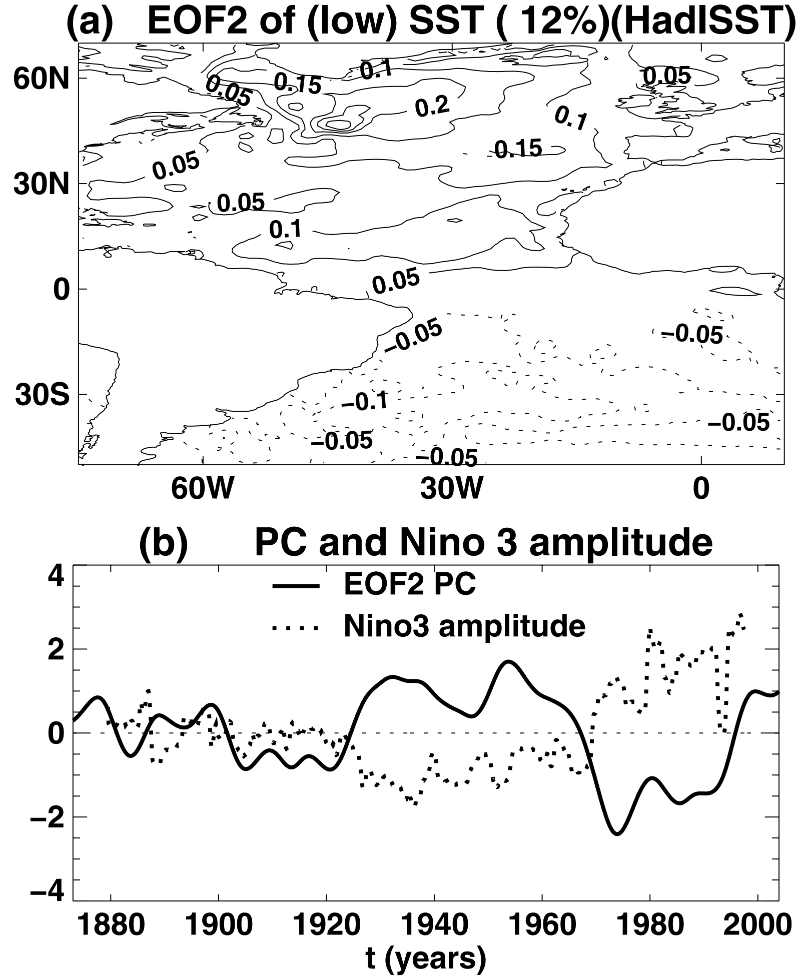

Figure 9: a) Spatial pattern of low frequency variability in the Atlantic representing the AMO, and b) its temporal variability. The hypothesis is that there is a link between the AMO and ENSO with weaker ENSO variability coinciding with the positive phase of the AMO. The link suggested here from data is supported by numerical experiments. From Dong et al. (2006).

there could be two stable states of ocean equilibrium, a vigorous thermally driven circulation like that observed now in the Atlantic with sinking at high latitudes driven by high latitude cooling of the surface water and another less obvious solution in which water sinks in the subtropics and rises at high latitudes. In this case the high density of the sinking water results from evaporation making the water salty and therefore dense. Perhaps counterintuitively it is this weaker haline-driven circulation which exists for the larger range of parameters. Stommel's model was very simple (and elegant) but not easily verifiable. There are various extensions to the basic paper nicely covered by Rooth. There have been indications from more comprehensive models, however, that the AMOC could substantially weaken, if not shut down completely and one of the concerns has been that global change could induce such a collapse. None of the climate models used in the latest IPCC assessment indicates such a drastic outcome in the next century but the consequences are sufficiently severe that a major study was initiated by NERC in 2004 called RAPID [3] to measure the AMOC and improve understanding of processes contributing to its variability and stability. This programme has now ended but a follow-on programme has begun called RAPID-WATCH. One possible outcome of RAPID-WATCH is operational monitoring of the AMOC.

Vellinga and Wood (2002) have considered the consequences of a major collapse of the AMOC. They did this by a one-off injection of fresh water into the northern Atlantic, equivalent to freshening the top 800 m of water by about 2psu. The AMOC collapses in the first few years reducing the northward heat transport and associated heat release to the atmosphere. The collapse is not just felt locally in the Atlantic: a cooling covers most of the northern hemisphere. Though locally this may be up to 8K, the overall cooling is more like 1K and 1-3K in the European sector. The southern hemisphere experiences warming peaking about 40 years after the injection. There are changes to the position of the ITCZ as it is moved southward, leading to marked changes in precipitation of up to 1m/year. The wet season over Brazil becomes wetter, the SW monsoon becomes weaker and the rainfall over Europe is reduced, probably as a result of the atmospheric cooling. Hazeleger (presentation at the RAPID conference in Cambridge 2008) has also studied the impact of a major reduction in the AMOC. He finds large gobal effects but a rather small impact over Europe. The Bjerkenes compensation is operative wherebye a large reduction in ocean heat transport is compensated substantially by an increase in atmospheric transport reducing the impact on the top of the atmosphere radiation. Although Atlantic SSTs drop substantially, the reduction in western European surface air temperature (SAT) is surprisingly small in his model simulations. This results from strong local cloud feedbacks enhancing the cooling over the ocean but heating the continent.

The above experiments are transient experiments. The freshwater injection is at the start. The AMOC shuts down in a few years but as the coupled model runs free thereafter, the freshwater is free to adjust. Vellinga and Wood (2002) find that the model does not stay in the depressed state but that the AMOC recovers in about 80 years. Some other models which have been used to study 'water hosing' experiments do not recover; the AMOC remains suppressed.

3 Predictability studies

Several studies have been carried out to determine the degree to which decadal signals might be predicted. The strategy is generally to have a long run of a climate model. Some initial state is chosen and then a number of forecasts made from that date, differing in that some (small) perturbation is added to the initial condition. For some reason, probably convenience, usually only the atmospheric state is perturbed in the studies considered below. Almost always the ocean initial conditions are not altered. The ensemble will diverge, and the rate of divergence, compared to the size of the signal, gives an estimate of the time for which the signal might be

[3] see http://www.noc.soton.ac.uk/rapid/rapid.php

predicted. Usually the climate run uses fixed greenhouse gases and so gives an estimate of predictability of natural phenomena. The reforecast model is usually identical to the original climate model and so the estimates of predictability are under the assumption of perfect model and essentially perfect initial conditions. One would normally expect such estimates to be too optimistic. However, the models are not a perfect representation of reality and could be deficient in processes that could either increase or decrease predictability. Most models seem to have some variability in the AMOC for example, though the range and amplitude is quite variable as will be shown below. The correct level is not known although the RAPID-WATCH programme should go some way to quantify this.

The earliest study of decadal predictability was carried out by Griffies and Bryan (1997). They followed the above procedure of generating ensembles of forecasts from four initial states taken from a 1000-year climate control integration. One of the initial conditions gave rise to hindcasts which were best classified as following the Hasselmann (1976) integrated noise paradigm. The other three more closely represented damped oscillatory behaviour. The predictability horizon depended on the initial condition and on the variable chosen. They considered several: North Atlantic dynamic topography, SST, SSS, temperature at 170m (to filter out some of the noise in SST), as well as a measure of the AMOC and East Greenland SST. The least predictable variable was SST with a predictability horizon of about 2 years. The others were all in the range 6-20 years, depending on the initial state. They drew attention to salinity variations emanating from the GIN sea, hence the choice of SST in the East Greenland sea as a predictand. The Greenland Sea SST indicated predictability for 10 to 15 years associated with the strong fresh water anomaly there. See Dickson et al. (1988 and 2002) and Hack et al. (2003) for a discussion of observed variability in salinity.

In the late nineties, there was a major EU project PREDICATE (Sutton et al., 2003), which attempted to understand mechanisms for decadal variability and to assess predictability. One approach, using four AGCMs forced with observed SSTs showed that there might be some extended predictability especially for summer conditions. This study was rather different to that of Griffies and Bryan in that only AGCMs were used compared to Griffies and Bryan (1997) who used a coupled model.

Collins and Sinha (2003) performed a set of three 9-member ensemble hindcasts for start dates corresponding to strong, average and weak AMOC initial conditions. The initial conditions were taken from a long control integration of the HadCM3 coupled model. They found the AMOC to be predictable to 10 years or more and in a subsequent limited experiment even as long as 50 years. To consider more practical predictors they consider SAT over an area of western Europe (15°W-10°E, 35°N -60°N). The climate pdf is obtained from the long control run shown in Fig 10 along with the pdfs of the various hindcasts. The pdfs are obtained by assuming a Gaussian distribution and fitting the 9-member ensemble to it. The results are considerably more optimistic than those of Griffies and Bryan (1997) and Pohlmann et al. (2004) discussed below. However, the usual caveats apply; the results are based on a small ensemble and a limited set of initial conditions.

Pohlmann et al. (2004) considered the potential predictability of the AMOC using the ECHAM model. They found that the highest potential predictability is over the North Atlantic, Nordic seas and Southern Ocean. Over land they find little predictability except for some regions of the UK, Ireland, and that part of Iberia close to the ocean. There is little predictability of sea level pressure. They also found that North Pacific SSTs are much less predictable than North Atlantic SSTs. In contrast to the low estimates of predictability of atmospheric fields such as sea level pressure or surface air temperature over land, there is considerable predictability of the AMOC. In their model the AMOC is predictable to more than 10 years ahead, as is the north Atlantic SST and SAT over the North Atlantic. There is a seasonality to the predictability of SAT with winter being more predictable than summer. Predictability of the Drake Passage Throughflow is limited to about 5 years. These results are somewhat different to an earlier predictability study using the Hamburg LSG model (Grötzner et al., 1999). To partially address model dependency of results, Collins et al. (2006) considered a multi-model approach. Four coupled models were run freely for hundreds of years. A total of 3100 years was used in bringing models

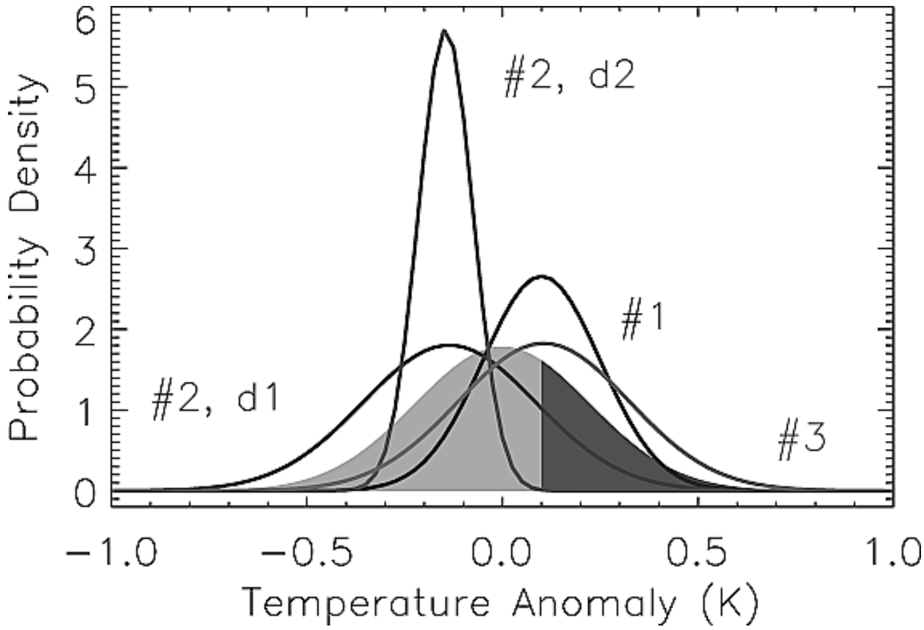

Figure 10: The probability distributions of decadal mean surface air temperatures (SAT) averaged over land points only, in the region 10°E-15°W, 35°N to 60°N. The grey shading indicates the climatological pdf with the warm tercile shown in darker grey. Also shown are the pdfs from 4 experients to test the predictability of SAT. Two pdfs have the same ensemble width as the control but are shifted, one to warmer conditions, one to cooler. One distribution is much narrower than the climatological pdf and displaced to cooler conditions, and the fourth is slightly narrower with the peak displaced to slightly warmer conditions. From Collins and Sinha (2003).

to statistical equilibrium and assessing their natural variability. A further 1340 years was used in making predictability experiments. Four models were involved in this PREDICATE project: HadCM3, ECHAM-OM1, ARPEGE3-ORCA2, and ARPEGE3-MICOM (BCM - Bergen Climate Model). The model ensembles were generated by perturbing the atmospheric state but no perturbations were added to the ocean state. There was quite a range in the amplitude and period of decadal variability. The most active was the ECHAM-OM1 and the least active ARPEGE3, but all four models did have variability. Predictability experiments were carried out as follows: each group chose a time when the AMOC was strong and a time when it was weak. Ensembles of experiments spanning at least 20 years were then carried out from these two types of initial state. Some models were additionally run for initial conditions in which the AMOC was in a more neutral state, but the results in this category were too few to draw meaningful conclusions.

Anomaly correlations and rms error curves based on these experiments suggest predictability of the AMOC out to a few years. A very favourable interpretation might be that the AMOC is predictable out to 10 years for strong conditions but only to 3 years if it starts in weak conditions. Predictability for SAT is considerably less, even though the SAT is defined over the region 40°-60°N, 50°-10°W, i.e. a predominantly ocean region. Another rather less optimistic result is the relationship between decadal-mean AMOC and decadal-mean SAT. There is a general sense that the northern hemisphere is warmer when the AMOC is stronger and transporting more heat poleward but the agreement over land areas such as Europe is weak. There is a general impression that the tropical ocean and southern hemisphere are cool when the AMOC is strong. This is especially marked in the case of the ECHAM-OM1 model which has a particularly strong AMOC variability. Presumably the differences in the levels of statistical significance result from the different levels of signal versus noise in the models but the mechanisms giving rise to decadal variability may well be different in the models. At the current level of knowledge it is not possible to say one model is better than the others and or that it gives a more reliable estimate of potential predictability.

As mentioned earlier, a major experiment is in progress to measure the AMOC at 26.5°N, a convenient latitude as the western boundary current is reasonable well monitored at this latitude and a few earlier estimates of mass and heat transport have been made as far back as 1957 (Bryden et al., 2005). This latitude is close to that at which the ocean heat transport in the Atlantic is maximum though the mass transport is maximum at somewhat higher latitude. This experiment was denoted RAPID in its initial phase and RAPID-WATCH in the follow-on phase. One motivation of establishing and maintaining the observing array is to provide early warning of a potential rapid collapse in the AMOC and the potential impact on European and North American climate. There is evidence of rapid changes in the AMOC from paleo data (Bond et al., 1997) and numerical models show rapid changes in the AMOC in response to imposed fresh water changes in the Nordic seas. However, long integrations of coupled models with normal climate conditions also show rapid, though not extreme changes, such as a collapse or reversal of the AMOC. Hawkins and Sutton (2008) consider rapid changes in the AMOC in a long (more than 1000 year) control intgration of the HadCM3 model and consider the predictability of such events. They find changes in T and S in the Nordic seas precede changes in the outflow of dense water through the Denmark Straights overflow, which in turn precede changes in the AMOC by several years. Anomalies in the AMOC and heat transport then lead to changes in European and North American climate several years later. Because these events seem to be triggered by changes in surface salinity and temperature in the Nordic Seas leading to changes in convection and flow through Denmark Straights, they advocate additional measurements in this region in order to assist in model prediction.

The changes in salinity which seems to lead changes in T by a couple of years or so, might be related to NAO changes but this is unclear. The physical connection between Denmark Staights Overflow AMOC and heat transport suggests predictability after the event has been triggered even if the trigger itself is not particularly predictable. One should note, however, that the model used for these studies has some weaknesses; in particular the Denmark Straights has been widened/deepened to increase the amount of Throughflow. An interesting feature of the SST anomalies found in response to changes in the AMOC is that the cross equatorial dipole is only weakly present.

Recently, Hawkins and Sutton (2009) have considered decadal predictability by constructing a simplified version of a coupled general circulation model. Based on 1100 years of coupled integration of the Met Office HadCM3 model, they constructed a linear inverse model (LIM) and verified that this was a fair representation of the full nonlinear coupled GCM. Forecasts of SST and the AMOC using this model indicated predictability to considerably beyond 10 years. An advantage of using the LIM approach is that it is relatively easy to calculate the rapidly growing patterns (optimal perturbations). These suggest that a particularly sensitive region is the Nordic Seas and that enhanced observation of this region might be useful for improved initialisation of decadal forecasts.

Tzipermann et al. (2008) have also considered patterns of rapid growth using LIM but training the LIM model on the GFDL CM2.1 coupled model. Interestingly, they too find forecasts sensitive to errors in the Nordic Seas, particularly the Labrador Sea. The GFDL model indicates more rapid growth of errors than HadCM3 and therefore the predictability limits are substantially shorter. This reminds us that the concept of predictability limits is model dependent. Perhaps as models develop, some convergence of the predictability horizons will result but at present the range is substantial.

To summarise this section, most models indicate some predictability of SSTs in the North Atlantic, North Pacific, and Southern Oceans, and some indication of predictability in the tropical Pacific. The degree of predictability is variable, probably being both model and initial condition dependent. The degree to which conditions over land are predictable is highly model and possibly initial condition dependent. Some models e.g. Collins and Sinha (2003) and Hawkins and Sutton (2009), give optimistic results, others such as Griffies

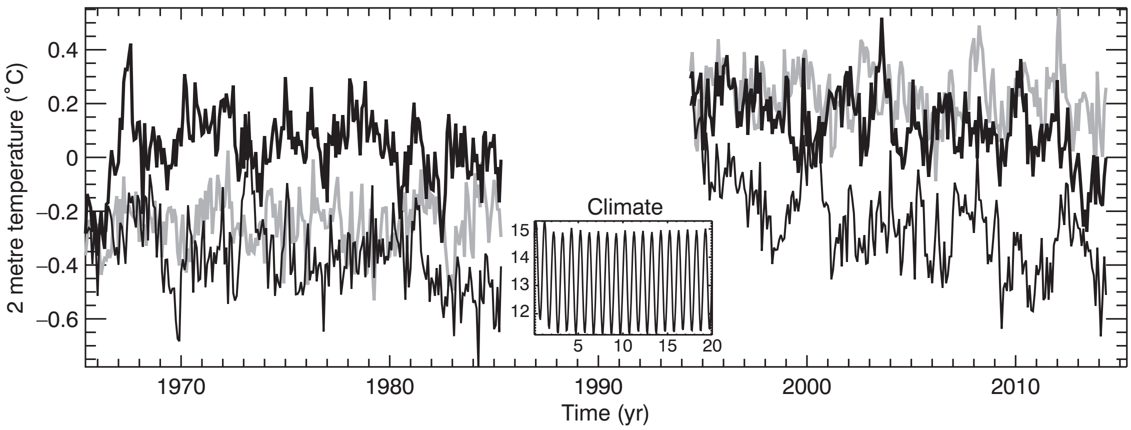

Figure 11: Global mean near-surface temperatures (T2m) plotted from the two start dates. The control is in light grey, runs with 1965 levels of GHGs in thin black and 1994 levels in thick black. The central panel shows the model climate. From Troccoli and Palmer (2007).

and Bryan (1997), Pohlmann et al. (2004) and Tzipermann et al. (2008) indicate shorter predictability horizons. However, the ensemble size and number of initial conditions used are very limited and so any conclusions must be treated with caution. The model resolution is generally poor adding to the uncertainty. The metrics used to assess predictability are rather variable making it difficult to intercompare results between models. Even basic quantities such as the AMOC are defined differently by different modelling groups and even by the same group at different times [4], but it seems to be universally the case that ocean quantities such as the AMOC are more predictable than atmospheric quantities. Rapid changes in the AMOC are also predictable on timescales of years, at least after the event has been initiated in the high latitude Atlantic. Whether the onset of such an event is also predictable is less clear.

3.1 Natural Variability vs forced global change

There are two classes of climate prediction: the initial value problem of prediction from a given realistic state and the change in climate in response to imposed external forcing such as greenhouse gases. Lorenz (1975) called these predictions of the first and second kind. In practice any forecast for a few years to a few decades is a mixture of both. In the previous we have mainly considered potential predictability of the first kind. But any realistic forecast from current conditions must take into account the role of greenhouse gases. Models can be used to determine the timescales for which the initial conditions are important before the effects of increasing greenhouse gases take over.

This has been considered by Collins and Allen (2002), Latif et al. (2004) and more recently by Troccoli and Palmer (2007). A schematic of the relative uncertainty from ocean inital conditions and for uncertain greenhouse gas concentrations is shown in Cox and Stephenson (2007). In no case is there a comprehensive set of experiments. Collins and Allen (2002) used an opportunistic set of two climate change integrations and a control integration. Hindcasts from only one start date were used (September 1974). Nonetheless some potentially interesting results emerged. It takes about 10 years for the global mean temperature in the greenhouse-gas run to be distinguishable from the control integration. Northern hemisphere land temperature is distinguishable on roughly the same timescale but it takes much longer before the climate change signal in north Atlantic SST is apparent. They argue that the global signal is detectable earlier than a regional signal because there is more

[4] This is important because the simulation of the AMOC varies considerably in models. In the ocean analysis ORA-S3, the AMOC is maximum at 48°N, whereas in some coupled models the maximum may be at as low as 30°N. The RAPID array of measurements is at 26.5°N where the heat transport is near its maximum though the mass transport is not.

filtering of atmospheric noise in the larger domain. The Northern Hemisphere land signal is more detectable in winter than in summer. For predictability of the first kind they estimate global temperature might be predictable up to two years in advance, North Atlantic SST up to a decade ahead but there is little predictability in land temperature. They do not perform many experiments, only one start date and only a small ensemble size. Although prediction skill may be related to ensemble spread, this was not shown. It has not been easy to show such a relationship in predicting shorter climate signals such as ENSO: so perhaps it will be difficult in the case of decadal prediction also.

Troccoli and Palmer (2007) used a coupled model consisting of the ECMWF atmospheric model (cy28r1) coupled to the OPA ocean in its ORCA2 configuration. The resolution is about 2 degrees in both directions, except in the equatorial region where the meridional resolution in the ocean is enhanced to allow better representation of equatorial waves. Two start dates were chosen (May 1965 and May 1994). For each start date an ensemble of 4 hindcasts was made for 20 years. Realistic greenhouse gas concentrations were used. In this limited case of just two start dates, only a binary validation of the forecasts can be made: did the model correctly predict colder or warmer conditions. To assess the sensitivity of the model to changing greenhouse gases, two additional hindcasts were made in which greenhouse gas concentration appropriate to 1965 were used both for integrations started in 1965 and in 1994. Similarly, the 1994 initial conditions were used with both 1965 and 1994 concentrations. From the 4 x 20-year integrations one can get some sense of the speed of adjustment to changing greenhouse gases. Fig 11 shows the results of these 4 experiments. One can clearly distinguish the integrations with high and low greenhouse gas concentrations. (The CO_2 concentrations were 320 and 359 ppmv). Troccoli and Palmer (2007) estimate that the greenhouse gas effect overwhelms the influence of initial conditions in 5 to 10 years. The quantity examined is the global mean temperature. The impact on the AMOC is not known. However, the set of integrations is small (only 4 members) and the 1994 concentration when applied to 1965 initial conditions implies a jump increase in gases at the start of the integration (by 39ppmv). Likewise in the 1994 integration there is an abrupt drop. So these results need further refinement but the suggested timescales for surface quantities are plausible, and broadly consistent with earlier estimates of Collins and Allen (2002).

The impact of greenhouse gases on the AMOC was also considered by Latif et al. (2004). Four different greenhouse gas simulations with CO_2 increasing at 1% per year were started from four different points in a long control integration with fixed greenhouse gases. Latif et al. (2004) argue that the global mean temperature increased rather monotonically and was not sensitive to the initial conditions. By contrast the AMOC follows the control run for several decades before it is influenced by the increasing GHGs i.e. the evolution of the AMOC depends strongly on the initial conditions. They use the ECHAM5/MPI-OM coupled model (closely related to the ECHAM-OM1 referred to earlier) which has very strong multidecadal variability. In the following section results of prediction experiments are presented. These reforecasts using initialised ocean conditions are frequently compared with simulations in which no initialisation is performed. As the GHG forcing is the same in both, the importance of the initial conditions can be assessed. Knight (2009) has recently tried to separate natural variability in the AMO from the anthropogenically forced response using a suite of model (CMIP) simulations with both natural and anthropogenic forcing. He concludes that the North Atlantic is warming faster than expected from anthropogenic forcing and that the AMO entered a positive phase in the 1990s, which could last for up to three decades.

4 Prediction experiments

Models which have been used for predictability studies could be used for prediction experiments as well, provided a mechanism for initialising the model can be developed. External forcing must be taken into account and so predictions of greenhouse gases and solar activity should be included. In addition to real-time predictions, a series of hindcasts, sometimes called reforecasts, is needed. For these, past knowledge of greenhouse gases, aerosols and solar activity is used. Partial knowledge of volcanic activity is also included in the sense that any knowledge of volcanic aerosols prior to the start of the hindcast is used in the initial conditions. During the hindcasts the volcanic aerosol decays with a typical decay time of 1 year. Although it is likely that climate overall will warm this century, the details of the warming are likely to be influenced by the state of the climate system now, especially the state of the ocean, with natural variability likely to have a strong influence on the climate of the coming decade. Climate prediction experiments should therefore start from realistic initial conditions of the ocean, cryosphere, and land system.

The most advanced decadal forecasting system currently used operationally is that of Smith et al. (2007), called DePreSys (Decadal Prediction System). This is based on HadCM3 as in Collins et al. (2006). It uses an anomaly initialisation scheme, meaning that the model is brought to statistical equilibrium before being initialised. An ocean analysis system is run, assimilating data and creating first an ocean analysis and then, by subtracting the ocean climate, an ocean anomaly. This three dimensional anomaly field is added to the ocean state of the coupled model to initialise the ocean component of the coupled model. Similarly atmospheric anomalies from the ERA40 or ECMWF operational atmospheric analysis are added to the coupled model atmospheric state. The way the ocean analyses are produced is described in Smith and Murphy (2007). A number of four-member ensemble hindcasts have been generated out to 10 years from four start times per year (1 Mar, 1 June, 1 Sept and 1 Dec), from 1982 to 2001. An additional set of hindcasts was made from a simulation of the 20th century using observed levels of GHGs but not starting from the observed state of the atmosphere or ocean. Comparison of the two experiments, denoted as 'DePreSys' and 'uninitialised' by Smith et al. (2007) allows an assessment of the importance of initial conditions on the skill of the hindcasts. One measure of skill used is rmse (root mean square error) of global mean surface temperature. Results show that the error is higher in the uninitialised case than in the hindcasts initialised from 'realistic' initial conditions. They argue that the improved skill in the DePreSys case results from the better prediction of ENSO, but this effect is just for the first year or so. The improvements in subsequent years is linked to improved heat content initialisation. No comparison is made of the different responses of AMOC in the two sets of integrations. The above results assess the skill in DePreSys based on the relatively short 20 year period from 1982. A real-time forecast is made from June 2005, based on a 20-member ensemble out to 20 years. In their experiments internal variability leads to cooler conditions in the first few years after 2005, offsetting the effects of anthropogenic forcing. As a result, their initialised forecasts indicate no net warming until after 2008.

Fig 12 shows the impact of starting the forecasts from initialised conditions by comparing the rms error of the near surface temperature from DePreSys forecasts with the rms error from the uninitialised. Panel A shows the rms error of 9-year hindcasts from the uninitialised hindcasts and B) from the DePreSys (i.e. initialised) hindcasts. Panel C) shows the difference between A and B. Panel D shows the difference in rms error in the predictions of upper ocean heat content. There is a strong correlation between panels C and D. The disappointing aspect of panels C and D is that the skill over the North Atlantic is lower in the initialised case. On the positive side, the skill is higher in the Indian ocean and the southern oceans in DePreSys. We will return to this point later.

DePreSys is expensive in terms of the number of years of computing; a total of 6400 years of coupled integration went into the hindcasts, and 400 into the forecasts. To this must be added 1300 to bring the model into equilibrium and 320 years used for forecasts from uninitialised initial conditions, making a total of 8520 years. (In fact the total is slightly more than this as the ensemble members for the 2005 forecast includes integrations started earlier than 2005 as well as those from 2005.)

Smith et al. (2007) show that the spread in the forecasts and hindcasts of DePreSys is too small compared to the growth of error. This can be improved by combining forecasts from several models (as will happen in ENSEMBLES) or by representing more uncertainty associated with parameterisation, as can be done by perturbing the values of 'free' parameters within parameterisation schemes. Such an extension to DePreSys

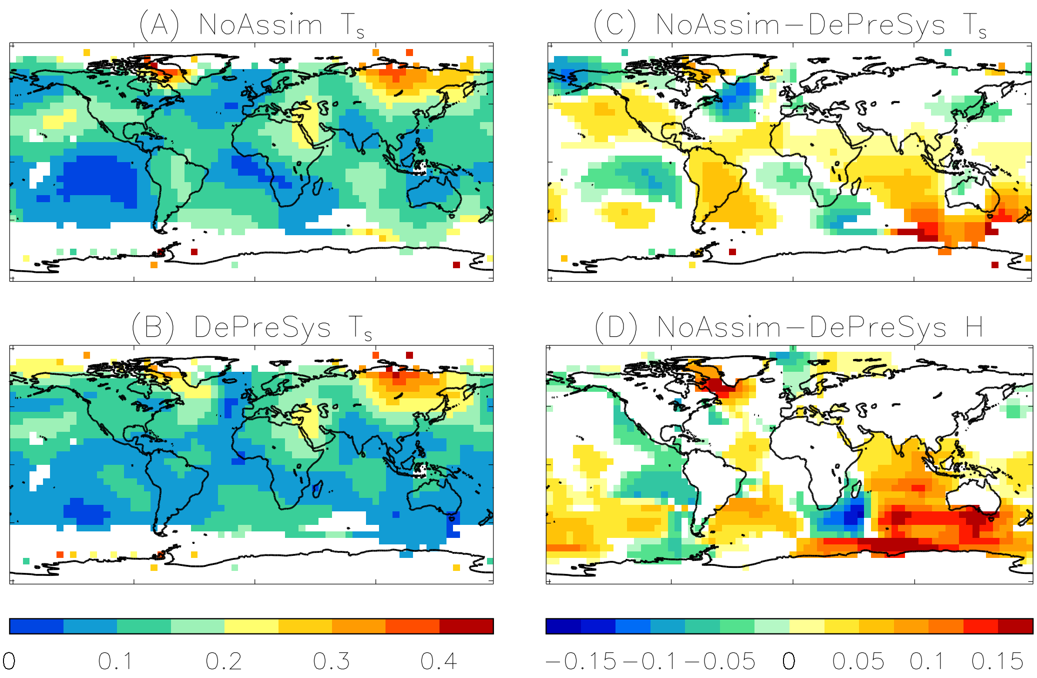

Figure 12: Impact of initial conditions on forecast skill. A and B indicate rms error of 9-year mean hindcasts of near-surface temperatures for the uninitialised and DePreSys hindcasts initialised over the period 1979-2001, and C shows the differences where they are significant. D is similar to C but for upper ocean heat content. From Smith et al. (2007).

has been developed for the EU project ENSEMBLES. This is called DePreSys-PP (PP for Perturbed Physics). In this scheme, various parameters or combination of parameters have their values changed commensurate with the perceived uncertainty in these parameters. This approach has some interesting features but also some drawbacks. Considerable extra computation is required. Changing parameter values means that the model mean state will change. Further, individual members could have quite large local systematic errors even though the global mean temperatures were similar. To overcome this difficulty, the PP approach includes a flux correction. In a prototype experiment (Collins et al., 2006) the flux correction was calculated for each ensemble member by running the coupled model for approximately 300 years until it came into statistical equilibrium. During this phase a strong relaxation was applied to observed SST and SSS. These correction terms were averaged over the last 50 years and then added to the model as a flux rather than a feedback term. However, although a nearly equilibrium state of the model should be reached by this means, there is no guarantee that the state will be stable. In fact there is good reason to expect it not to be (Tzipermann et al., 1994), and that is what was found (Collins et al., 2006). However, if the feedback terms are reduced, especially on salinity, then a more stable equilibrium state can be found. So, for ENSEMBLES, the feedback terms on T and on S were reduced to give damping times of 30 days and 120 days respectively compared to the 15 days used in Collins et al. (2006). The model was then run with the feedback terms on for 100 years and in the flux adjustment phase for 150 years (Glenn Harris, private communication). Since this procedure is used for each ensemble member of which there are 9, it is quite costly requiring 2250 years of coupled model integration before the start of the hindcast integrations. The hindcasts and forecasts are also costly; 4050 years are needed for the 10-year forecasts and 1620 years for the 30-year forecasts, i.e. about 8000 years in total. Fig 13 shows the drift in the original coupled model (Gordon et al., 2000), in the model which had an unstable equilibrium (Collins et al., 2006) and in the revised flux-corrected case for ENSEMBLES. The systematic error in the latter case appears to be quite modest, especially compared to the interim system which had large systematic errors in the North

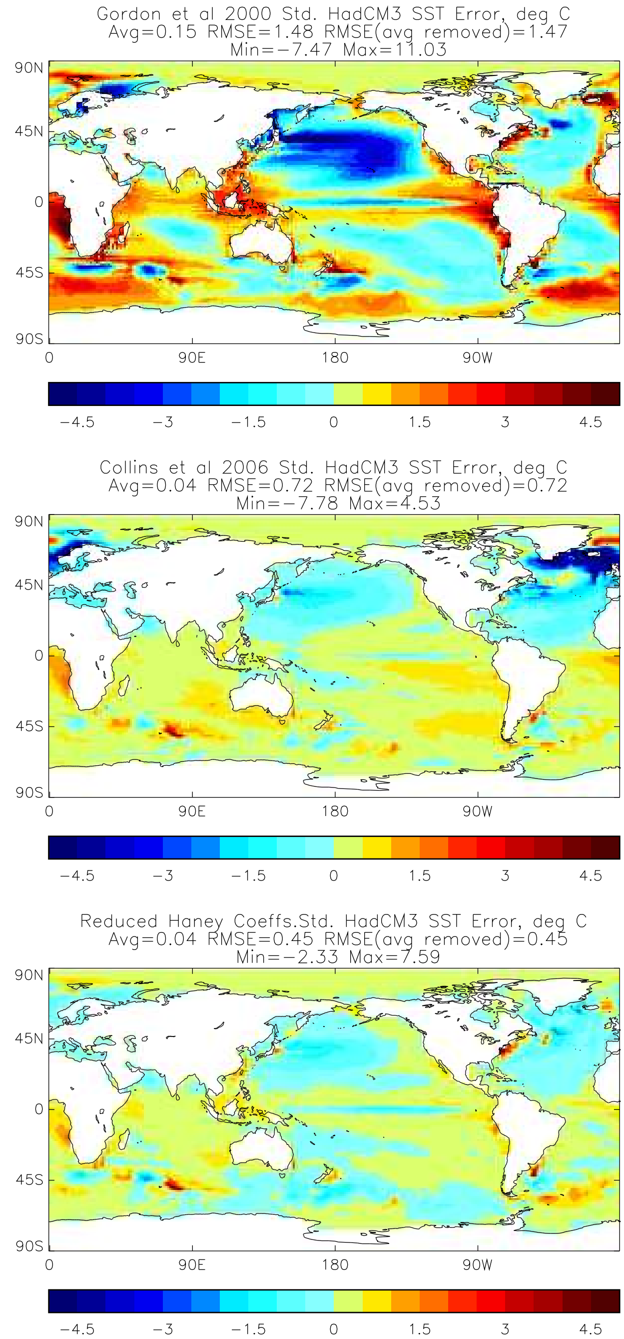

Figure 13: Drift patterns for a) HadCM3, b) the perturbed physics version of the HadCM3 model used by Collins et al. (2006), c) DePresSys-PP. From Collins, private communication.

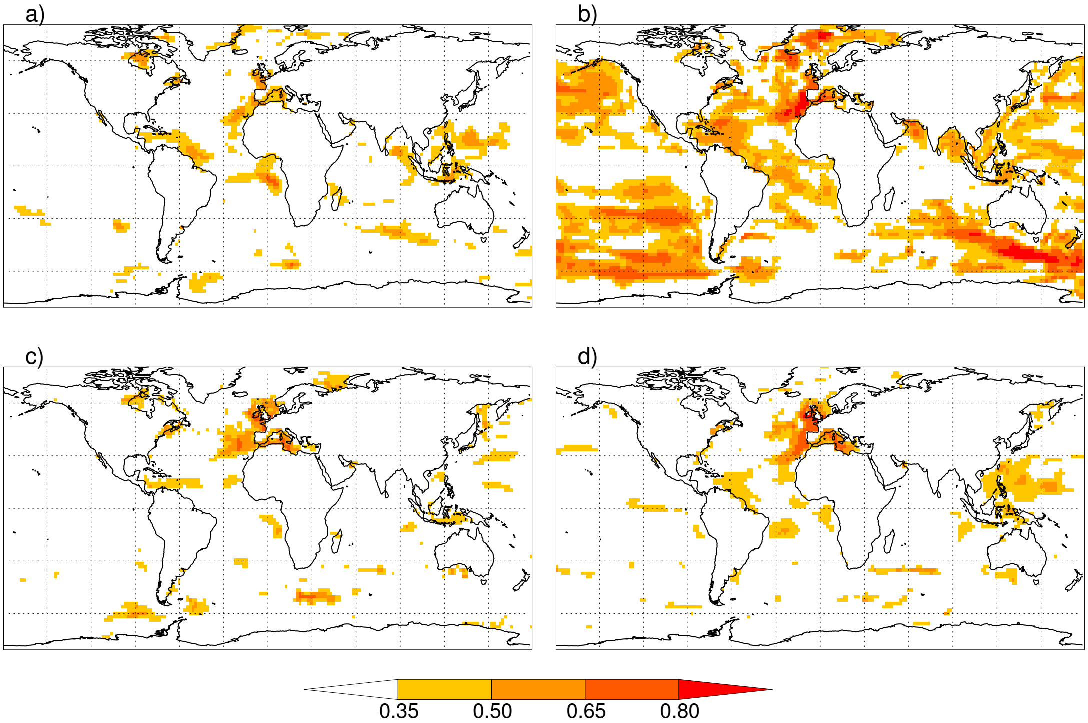

Figure 14: Anomaly correlation coefficient between the sea surface temperature of the GECCO synthesis and the ensemble mean 20C experiment (a) and the hindcasts for the first year (b), year 5 (c), and year 10 (d). The coloured areas are significant at the 95 percent level according to a t-test. From Pohlmann et al. (2009).

Atlantic, even though there was little drift in global mean temperature.

Pohlmann et al. (2009), have also performed a set of decadal hindcasts and forecasts using an anomaly approach similar to Smith et al. (2007), in that they run the coupled model to statistical equilibrium before initialising it. The approach differs from Smith et al. (2007) in the way the ocean and atmosphere are initialised; an ocean analysis based on GECCO is used to provide anomalies of the ocean state. The coupled model is relaxed to these ocean anomalies and the atmospheric state allowed to come into adjustment with the model SSTs. There is no feedback on SST in the ocean and so the surface layer temperatures need not be close to observed. Since the atmospheric state is determined solely by the model SST it would not be synoptically close to observed. Some aspects of the low frequency response between atmosphere and ocean might be in better balance as the SST is allowed to adjust in a coupled environment to some degree but other features such as the MJO would not be initialised close to reality. The extent to which initialisation of such processes is important in decadal forecasting is unclear.

A single 10-year forecast is made every year from 1951 to 2001. Hindcasts are mainly compared, not to observations, but to their analysis. Fig 14 shows results from the hindcasts based on sea surface temperature. Anomaly correlations for a lead of 1, 5 and 10 years are shown in panels b, c, and d. A control set of three integrations, made for the IPCC-AR4, can also be assessed. By comparing these with the initialised re-forecasts one can assess the importance of oceanic initial conditions. Results based on using the control as a hindcast are shown in panel a. As one would expect there is almost no skill in the control but some skill in the hindcast set (panel b). Interestingly, the skill in the tropical Pacific is poor, probably because the GECCO analyses do not well represent observed temperatures in the surface layers of the ocean. The 5-year forecasts show some skill particularly in the Nort-East Atlantic. Even at 10 years there is skill to the west of Iberia.

In addition to comparing the model-predicted surface air temperature to the analysed SAT, they compare the skill with other prediction methods. One is a simple statistical scheme - damped persistence using as a damping time the lag-1 autocorrelation coefficient from observations. The other is a trend forecast. The trend is calculated from the prior year, pentade or decade to the initial year, pentade or decade, and then used to extrapolate into the future. The skill of these two methods is lower than that of the coupled model in general but for shorter forecasts (shorter than 3 years) damped persistence works as well as the coupled model and for decadal averages the trend works as well as the coupled model. In the intervening periods, the coupled model is the best.

They also consider the predictability of the AMOC comparing the predicted values to the GECCO analyses at 48°N and find correlation skill of 0.81, 0.93 and 0.97 for annual, pentadal and decadal mean values for the initialised model but no skill for the control runs, indicating that the initial conditions of the ocean hindcasts contain useful information about the AMOC. Their results indicate that it can be predicted for a few years ahead and that it leads the North Atlantic averaged SSTs by about 5 years, presumably through the AMOC influencing ocean heat transport. The latitude used to measure the AMOC is 48°N, so not directly comparable to Keenlyside et al. (2008) who use 30°N (see later), nor to Balmaseda et al. (2007) who use 26°N, nor to 'observed' values from Cunningham et al. (2007) at 26°N. The analysed AMOC in GECCO is maximum at 48°N, but the coupled model if running freely has it located considerably further south. The assimilation analysis agrees with the GECCO analysis quite well at 48°N but not at all well at lower latitudes such as 26°N where the RAPID measurements are made. This seems somewhat strange as a rather strong relaxation is applied (10-days). The authors attribute the lack of agreement to Rossby wave activity. Forecasts of globally averaged SSTs are generally poor when compared to observed SSTs, thought to be because the SSTs in the GECCO analyses are poor, especially in the southern hemisphere. Pohlmann et al. (2009) go on to make a ten year prediction. However, as the GECCO analysis ends in 2002, their forecasts are started in 2002. Based on a 7-member ensemble they predict increased SAT over North America, the North Atlantic and Europe for the 6 years 2002-2007 relative to the radiatively forced integrations. There are no results given beyond this time. Since only a single 10-year hindcast is made every year and the control set of hindcasts come from previous 20C integrations, this system is vastly cheaper than the standard DePreSys or DePreSys-PP.

An alternative, even simpler strategy for initialising and running the forecasts has been adopted by Keenlyside et al. (2008). The assumption is that the atmosphere responds correctly to low frequency variability in the SST, and the resulting surface winds (and heat and fresh water fluxes), when used to force an ocean model, generate a reasonable representation of the ocean state. However, it is far from clear that the low frequency atmospheric response at middle latitude can be deduced from SST, and so the results seem somewhat controversial. Rodwell et al. (1999) forced an atmospheric GCM with observed SSTs for the period from 1870 to 1997. For the period 1947-1997, which they discuss in their paper, they were able to reproduce some of the NAO variability but the amplitude was weak. For the earlier period the results were not good. Bretherton and Battisti (2000) question the use of SST forced-experiments, as do many others. See for example, Palmer et al. (2008). On the other hand, Rodwell et al. (2004) have considered the impact of SSTs on the atmospheric response further. They have taken five atmospheric models and forced then in the same way using the same SST patterns. The patterns were extracted from observations to optimise the response. They vary seasonally but are repeated from year to year. They find considerable agreement between models in the response, including the NAO and further the response is in good agreement with that related observationally to the SSTs. An extension, but just using the HadAM3 model is to partition the response between tropical and extratropical SSTs. They find a considerable amount of the extratropical response is associated with SSTs in the Caribbean and tropical Atlantic, although that associated with local mid-latitude SSTs is not negligible. Based on these contradictory results, it is therefore unclear that the strategy of Keenlyside et al. (2008) will work. Since NAO forcing is an important ingredient in forcing the AMOC (Eden and Willebrand, 2001; Latif et al., 2004), it is by no means certain that useful forecasts will be possible using just SST information for initialisation. Nonetheless it is instructive to try. In fact they find that the NAO response in their SST-forced atmospheric analyses does not match the observed

NAO variability well.

As for Smith et al. (2007) and Pohlmann et al. (2009), this is an anomaly initialisation procedure as only SST anomalies are used. Since the results are based on a hindcast set of 3 ensemble members initialised every 5 years from 1955 to 2005 and integrated for 10 years, this system, like Pohlmann et al. (2009) is very cheap compared with DePreSys, requiring a total of only 330 years of coupled model integration compared with the thousands used by DePreSys and DePreSys-PP. The external radiative forcing is as used by Smith et al. (2007), and as specified by ENSEMBLES. Additional resources are needed to bring the model to statistical equilibrium and to prepare the initial conditions by forcing in the SST anomalies. A further set of experiments was used which were not initialised. These experiments, denoted control, derived from 20C-RF runs are not purpose-made but rather, as in Pohlmann et al. (2009), are integrations already made as part of the 20C simulations. Consequently they are not as clean as the control set used by Smith et al. (2007) in that observed volcanic aerosol is used throughout the integration, rather than just at the start of the hindcasts with exponential decay thereafter as an approximation to the subsequent evolution. Using anomaly correlation as a measure of skill, Fig 15 shows the skill based on the set of hindcasts and also that from the uninitialised set. In the North Atlantic the results are better in the initialised hindcasts than in the uninitialised case. However, in the southern Indian ocean, Southern Atlantic and tropical Atlantic, this is not so. The assimilation hindcast set has skill in the tropical Pacific not present in the uninitialised 20C runs. Globally it is not clear that there is increased skill in the initialised set, but in the North Atlantic, there appears to be. This is the region they concentrate on though it is unclear to what extent this result is statistically significant since the sample set is quite small. Fig 15 can be compared with Fig 12 from Smith et al. (2007), who had less skill in the Atlantic sector for initialised hindcasts than for their control.

One further comparison that could be made with Smith et al. (2007) is in terms of global mean temperature. Keenlyside et al. (2008) predict a reduced warming compared to the 20C (i.e. the uninitialised) runs; in their predictions, global temperature does not rise until after 2010. This result is similar to Smith et al. (2007) who also show that their initialised forecasts are cooler than their control integrations for the next decade and there is no rise for several years after the start of the forecasts in 2005. In the Keenlyside et al. (2008) case, this may result from a slowing of the AMOC. The values of the AMOC in the initialised case drop faster than those for the control. However, they start higher than the control and always remain higher. So the results do not appear totally consistent. The north Atlantic SST is cooler in the initialised case consistent with a declining AMOC, but not with the AMOC being higher than the control throughout the forecasts . See Fig 3 in Keenlyside et al. (2008). The low frequency variability in the MOC is large (10Sv peak to trough). As Smith et al. (2007) do not analyse the AMOC, we have no way of knowing what their model produces. As expected from the results summarised in a previous paragraph, they find that the simulated NAO forced by observed SST is poorly related to that observed. The predicted values are not given but there is no reason to expect the NAO to be well predicted if it can not be well simulated with observed SSTs.

Keenlyside et al. (2008) could investigate further the reliability of their system. If their approach is successful because they initialise the upper ocean heat content correctly, then one should be able to verify the hindcasts of upper ocean heat content against their analyses of this quantity. This has not been done but the null hypothesis would be that just supplying (mainly tropical) SST with no particular way of penetrating this information into or below the mixed layer would not work. Further, salinity is not corrected and since at high latitudes salinity can have an important effect on density, this would further limit the effectiveness of the approach. Finally a filter is applied such that no SST information is used polewards of 60° and only a limited amount between 30° and 60°. The onus is on the authors to show their approach works for the correct reason and is not just serendipitous. If it can be shown that their approach is an acceptable way of initialising the coupled system by just using SST then this is a powerful result and would allow verification of the decadal system over an extended period prior to 1958 since SST analyses go back to the 1870s.

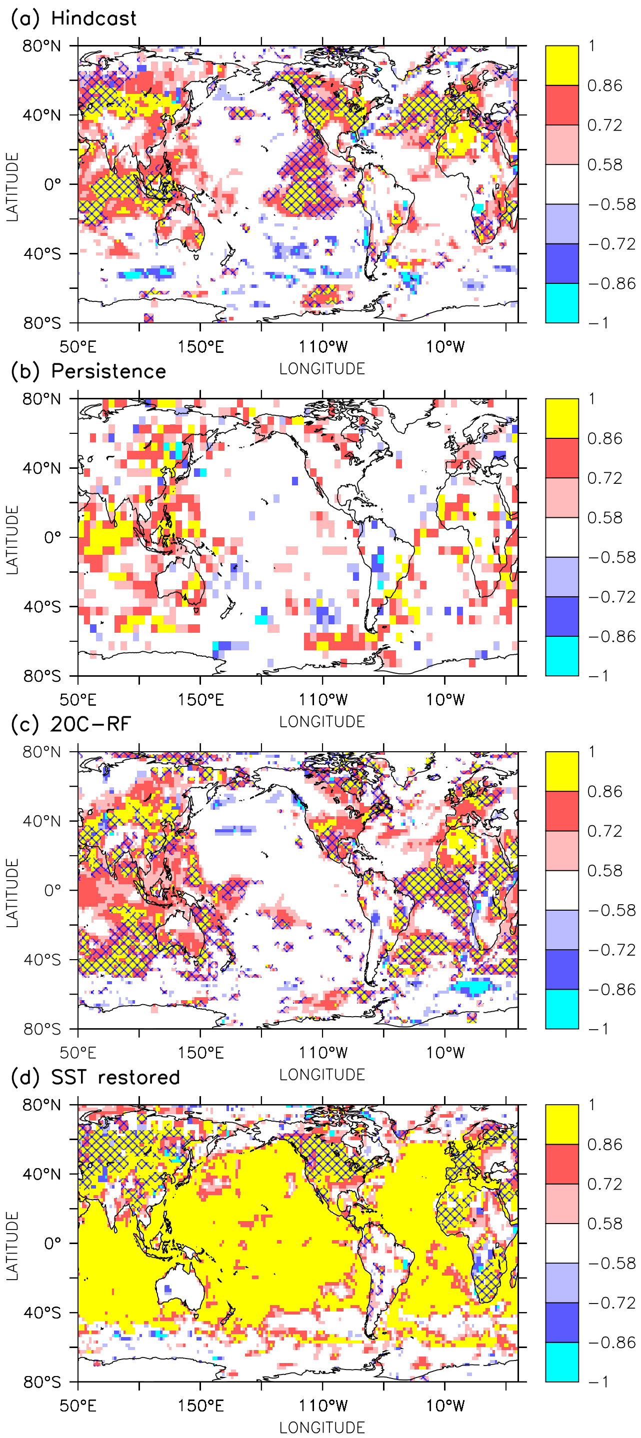

Figure 15: Plot of the correlation skill from the initialised and unitialised hindcasts. Areas where the skill of the initialised is statistically different from that of the control are shown hatched in a). Areas where the skill is worse are shown hatched on c). Taking the world as a whole the hatched area in a) is no greater than that in c). In the North Atlantic the skill of the initialised hindcasts is superior, but in the tropical Atlantic and the South Atlantic the opposite is true. From Keenlyside et al. (2008).

The approach of using just SST to initialise the coupled system for decadal forecasting has not been rigorously compared with other approaches using the same model in the context of decadal forecasting, but it has been compared for seasonal forecasting. Seasonal forecasting is a rather different problem to decadal forecasting as it emphasises the tropics over higher latitudes and the low latitudes may respond more strongly to imposed SST anomalies but Palmer et al. (2008) have argued that the experience with seasonal forecasting can give guidance to longer range forecasts such as decadal. In that spirit we record some results from seasonal forecasting here.

Keenlyside et al. (2005) and Luo et al. (2005, 2008) have both used the SST approach to make seasonal hindcasts and forecasts. The results appear to be model dependent. Keenlyside et al. (2005) found that the skill as measured by rms error and anomaly correlation in the key Niño indices was noticeably worse than the skill of other models he compared against (basically models participating in the EU project DEMETER). However this could result from having a poor forecast model and not necessarily a deficiency in the initialisation strategy. Luo et al. (2005, 2008) on the other hand claim the strategy of just using SST works well. To overcome this model dependence Balmaseda and Anderson (2009) intercompared the SST initialisation approach with that used in the ECMWF operational seasonal forecasting system. The same model was used in both; all that differed was the way the coupled model was initialised. In the operational approach full atmosphere and ocean analyses using all available data are performed independently. This gives a realistic analysis of both the atmosphere and ocean, though possibly also creates imbalances in the initial state leading to initialisation shock when the atmosphere and ocean are combined and the forecasts begun. In the SST approach, the atmosphere is forced with observed SST and the resulting surface fluxes are used to force the ocean. This method mimics the procedure used by Keenlyside et al. (2005), Luo et al. (2005, 2008) and Keenlyside et al. (2008). In all methods there is a strong relaxation to analysed SST fields from Reynolds OIv2 during the analysis phase. Using the same atmospheric component and the same ocean model, hindcasts from the two strategies for initialisation have been intercompared.

The root mean square error as a function of forecast lead time for the two methods is shown in Fig 16 for the Niño3.4 region. The operational approach is clearly the best, and the SST approach considerably poorer. It argues for using ocean data as much as possible and for using the best winds available. It is clear from this figure that the method which contains the most realistic information provides the most accurate forecasts even though in principle the model is not in the most balanced state. This result holds not just for Niño3.4 but for all regions considered, with the exception of the equatorial Atlantic. The model used for this study, in line with most coupled models, has trouble in forecasting events in the equatorial Atlantic. See Balmaseda and Anderson (2009) for further discussion of the results. Results from a similar set of experiments using the Met Office operational seasonal forecasting model Glosea3, are consistent with those from ECMWF that the strategy of using only SST information leads to a marked degradation in the forecast skill (Anderson et al., 2008). How pertinent these results are to the decadal prediction problem is unknown. Nor is it known whether there is any advantage in using anomaly initialisation as opposed to full field initialisation. Even for seasonal forecasting timescales this is not known. The one study to look at this, using a somewhat simpler model than is typically used operationally, indicated no advantage to using anomaly initialisation (Galanti et al., 2003).

4.1 How well can the ocean initial conditions be created?

The previous results and those of Smith et al. (2007) based on global SST argue that one should use as much information as possible in deriving the initial state. But even using as much ocean data as possible, how well can one analyse the ocean state? Here we consider two possible indicators which are relevant for decadal problems: the upper ocean heat content and the strength of the AMOC. Fig 17 shows the AMOC at 25°N in the Atlantic as analysed by several ocean analyses. The observed values from Bryden et al. (2005) are marked by red stars. The Cunningham et al. (2007) value is not included on this plot but the value can be obtained from

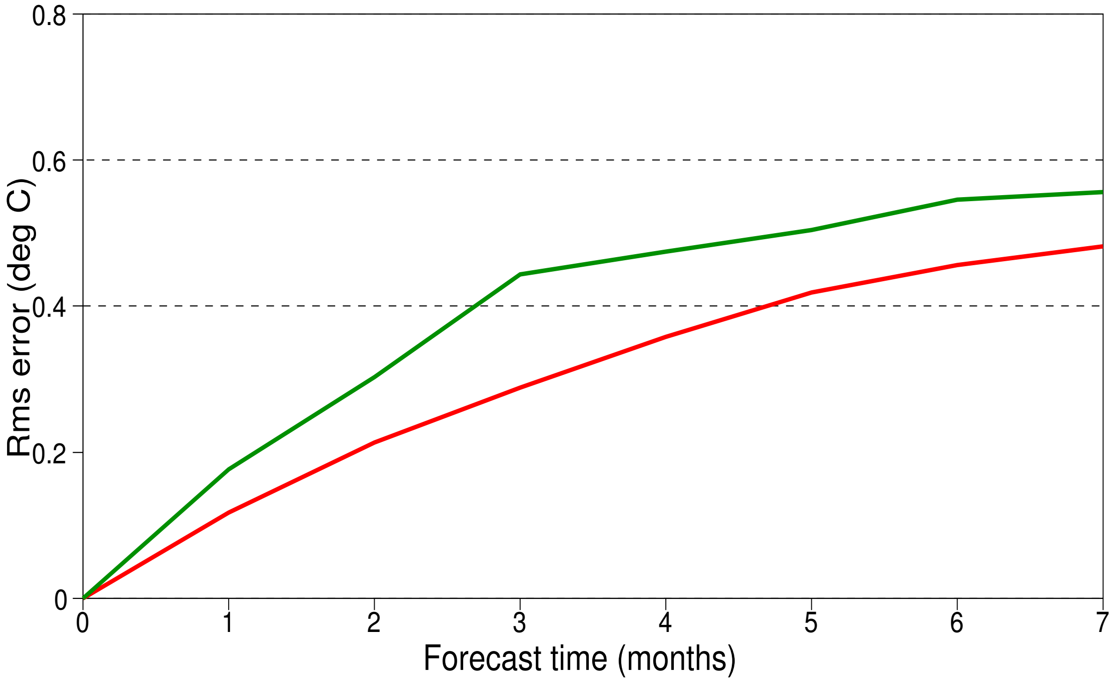

Figure 16: Plot of the rms error in the Niño 3.4 region as a function of forecast lead time for two sets of hindcasts. Green corresponds to the forecasts initialised using just SST, while red corresponds to hindcasts from initial conditions using all available oceanic and atmospheric data. The skill of the forecasts initialised just using SST data is significantly worse than the skill of those initialised using oceanic and atmospheric observations. From Balmaseda and Anderson (2009).

Fig 3. The results on Fig 17 are not from a controlled intercomparison set of analyses; rather they are *ad hoc* results from what various groups have been doing. The length of the analyses differ, the assimilation strategies vary, some use 3D-VAR, some 4D-VAR, some are OI-based, some analyse salinity, some not, the fluxes used to drive the ocean come from different reanalyses. Nonetheless the collection of results gives a fair assessment of the large disagreements there are between model analyses, reflecting the large uncertainty in estimates of the AMOC. To the curves shown in this figure, one could include those on Fig 3 from the ECMWF ORA-S3 ocean reanalysis. Based on these figures one might conclude that there is little hope of initialising the AMOC correctly. Further Pohlmann et al. (2009) indicate that they could not even reproduce the MOC at 25°N in their analysis. (Recall they nudge in the 3D analysis field from the GECCO analysis, which uses a different ocean model, and although they use strong nudging (10-day timescale), they say they can not reproduce the GECCO AMOC in their model.) They attribute this to Rossby waves, but the cause is not really clear. However, they make the point that there is better agreement in their model at 48°N than at 25°N. Fig 18 from Köhl (see footnote) shows the AMOC at this higher latitude. The agreement between the various model analyses in terms of temporal variability is not much better than that shown in Fig 17. There are some signs of hope, however. If one considers the values of the AMOC at 900m, at 48°N, (Fig 18) then the agreement appears better (the mean has been removed for each analysis separately). This figure covers only the years 1994-2002 when presumably there is greater agreement between the various atmospheric reanalyses in surface winds and fluxes. It is noted by Lee [5] that there is greater agreement with respect to the annual cycle than to the interannual anomaly. The choice of 48°N and 900m was made as being the location of the maximum in the MOC but this is not so in all models- for example see Fig 3 in Balmaseda et al. (2007), where the AMOC is maximum at 1200m.

Since Smith et al. (2007) indicate that upper ocean heat content is important for their forecasts we consider

[5] http://www.clivar.org/organisation/gsop/synthesis/synthesis.php

heat content from many ocean model analyses, presented by Balmaseda and Weaver at the GSOP meeting at ECMWF in August 2006. Balmaseda and Weaver intercompare the heat content in the upper 300m, 700m and 3000m across several models, globally, for the tropics, for the mid latitudes and for individual ocean basins. In Fig 20 the heat content for the North Atlantic and North Pacific are shown for the upper 300m. The agreement shown in this figure is poor. It is poor in the other regions considered too. See also Fig 10 of Carton and Santorelli (2008). The large spread in analyses is discouraging but one might take some solace from the fact that the spread present in the earlier period up to 1990 is reduced in the following years as the observing system improves. Unfortunately the spread increases again post 1998 although this might be due in part to a few outlier analyses. Although there is some scatter in the temporal variability, the various analyses generally represent the major peaks and troughs. One could perhaps argue that as the ocean observing system improves, new atmospheric reanalyses are generated and better assimilation schemes for the ocean analyses come on stream, the disagreement in heat content and AMOC might decrease. For example some analyses assimilate only T and neither analyse S nor adjust S to preserve the $T(S)$ relationship. Consequently the density field might be in substantial error. For example in the method used by Keenlyside et al. (2008), (their analyses are not included in the previous slides on AMOC and heat content), S was allowed to run free during the analysis stage and is thus unlikely to be realistic given the problems with precipitation in atmospheric models.

5 Some results from the ENSEMBLES project

The EU-funded ENSEMBLES project (www.ensembles-eu.org) has undertaken a decadal re-forecast experiment to assess the ability of state-of-the-art coupled forecast systems to predict interannual and decadal anomalies. ECMWF, the Institut für Meereskunde in Kiel (IfM), CERFACS and the UK Met Office all contribute to the experiment, the first three as part of a multi-model ensemble and the last one using the perturbed-parameter approach.

The common experimental setup consists of three-member ensembles initialized on the first of November once every five years over the period 1960 to 2005, with the first re-forecast starting in 1960. The simulations are run for 10 years. The ensembles start from either the best oceanic and atmospheric analyses available, either as absolute values (ECMWF and CERFACS) or as anomalies assimilated into the coupled model climate (Met Office), or from a set of coupled integrations nudged towards climatological model SSTs to which observed anomalies have been added (IfM). To generate the ensemble, wind and SST perturbations are used in the first two cases, while lagged initialization is employed by both the Met Office and IfM. All models employ observed greenhouse gas and sulphate aerosol concentrations and total solar radiance variations taken from observations or from an extrapolation of the most recent 11-yr cycle. Two of them (IfM and Met Office) also take into account the forcing from the volcanic aerosol present in the atmosphere at the initial time, damped with a time scale of one year. A summary of the initialisation strategies for the different model systems can be found in Weisheimer et al. (2007).

The Met Office system uses heat and momentum fluxes to prevent the coupled model from drifting too much from the model climate, as described in Section 4, whereas, ECMWF, CERFACS and IfM assume that non-linearities in the climate system do not significantly affect the drift of the model, which can then be removed *a posteriori* from the ensemble simulations. In addition to the basic experiment, the Met Office performs a 10-year re-forecast every year as well as a 30-year re-forecast every five years. This additional sample should provide more robust estimates of forecast quality and a near-term forecast of climate change.

The ECMWF approach described above is the same as that used in its seasonal forecast system viz to perform the best analysis of the atmosphere and ocean independently, assimilating as much data as possible. Any imbalance in the initial state and model drift resulting from model errors are removed *a posteriori*. By contrast,

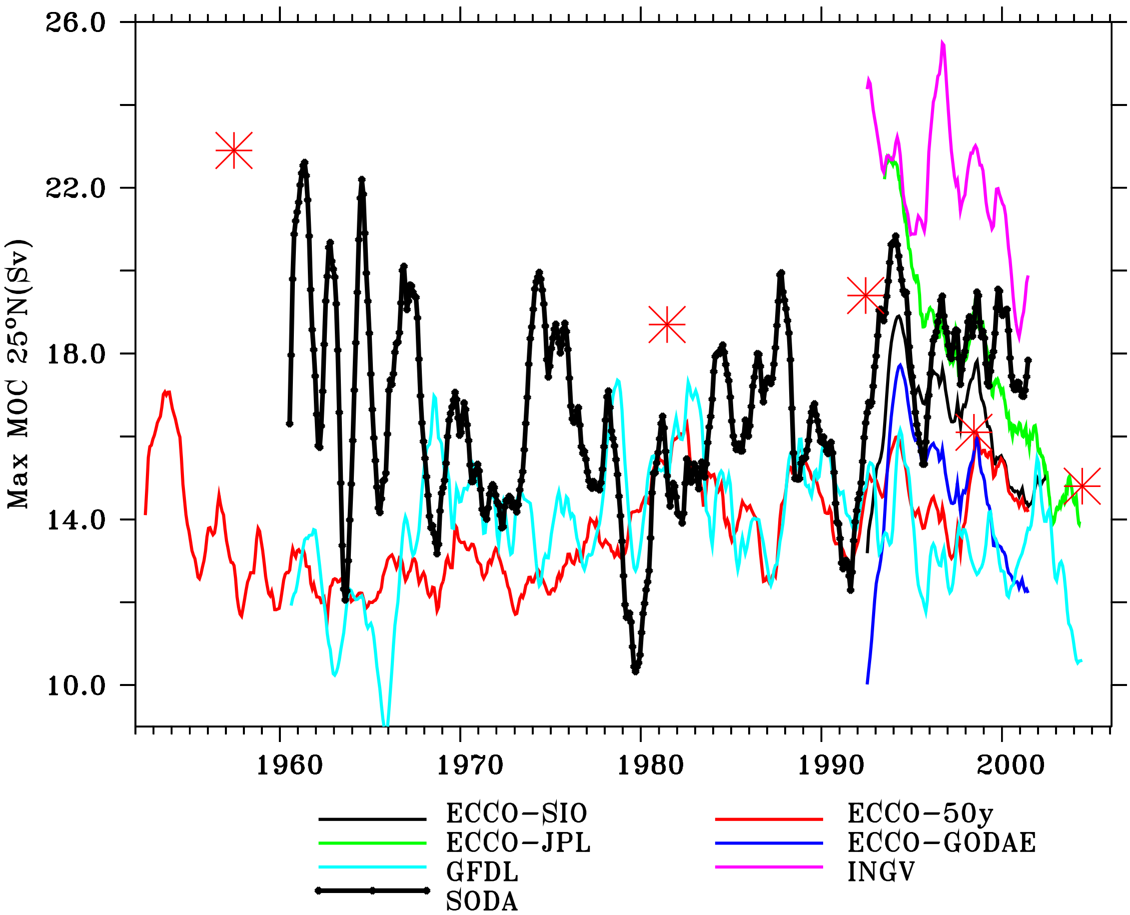

Figure 17: Temporal variability of the maximum strength of the AMOC at 25°N from several models. The red starts indicate data from Bryden et al. (2005). This figure should be viewed in conjunction with Fig 3. There is little agreement in the strength of the AMOC between models. Based on results from the GSOP meeting at ECMWF, August 2006. From Köhl, CLIVAR GSOP web site - see earlier footnote for http address.

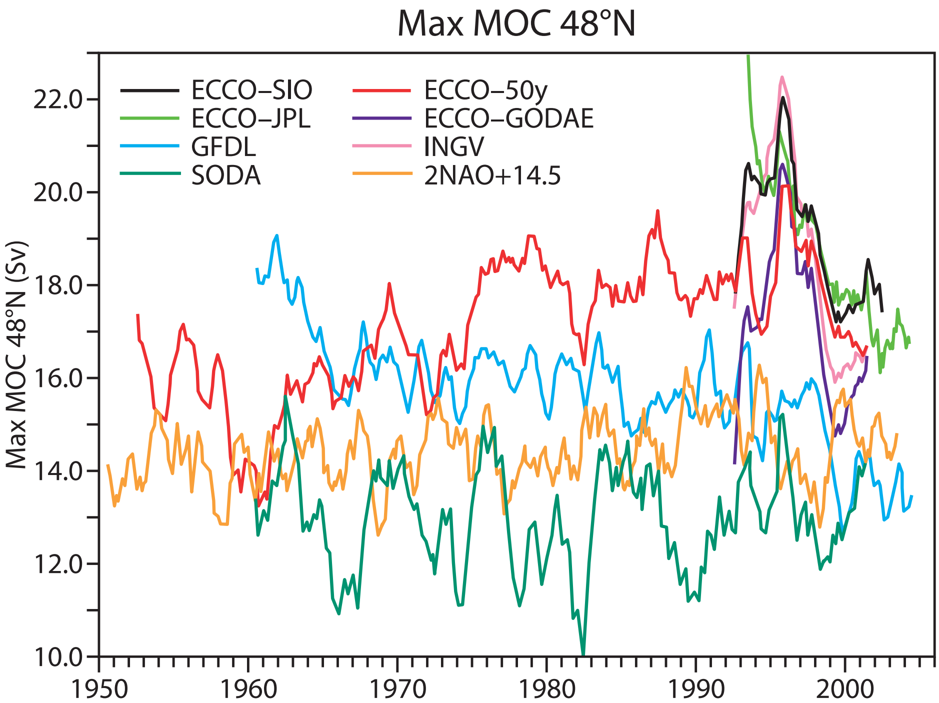

Figure 18: Plot of the AMOC at 48°N from various models. Based on results from the GSOP meeting at ECMWF, August 2006. From Lee, CLIVAR GSOP web site - see earlier footnote for http address.

the other three methods reviewed viz those of Smith et al. (2007), Pohlmann et al. (2009) and Keenlyside et al. (2008) use anomaly initialisation. As mentioned earlier it is as yet unclear if one strategy is better than another. In the seasonal forecasting context, Doblas-Reyes et al. (2009) found that seasonal predictions from DePreSys-PP have no advantage over a multi-model of the same ensemble size, although these results are not conclusive because the forecasting systems are different.

The assessment of the ENSEMBLES decadal re-forecasts is still on-going, but some preliminary results are given here. Anomalies of global-mean two-metre temperature for the three-member ensemble from the ECMWF system are shown in Figure 21. The predictions closely follow the time series from the analyses in spite of the small ensemble spread. The most recent forecasts suggest a small reduction of the warming trend. These results agree quite well with those obtained by DePreSys-PP and IfM.

Before we compare forecasts for the North Atlantic, we consider the different initial conditions for the AMOC. The maximum transport is shown in Fig 22 for five different model analysis systems. In four of these models the same atmospheric forcing is used. The fifth corresponds to the analyses generated solely from SST anomalies as used by Keenleyside et al. (2008) and discussed earlier. The weakest AMOC is that from the CERFACS 3D-VAR analysis using the NEMO 1° model with equatorial refinement. Not only is the mean weaker, but so too is the variability. Considerable variability is shown by the ECMWF analyses, which indicate a weak reduction in the AMOC in later years as discussed by Balmaseda et al. (2007). The Met Office AMOC is higher towards the latter part of the record than it is in the ealier half, but the most spectacular variability is that shown by IfM. Somewhere about 1985 there is a huge increase in the strength of the AMOC, followed by a rapid downward trend. This behaviour is intrigingly shown by all three ensemble members but not confirmed by the other analyses.

Not only is the mean strength of the AMOC wildly different between models but there also seems to be little

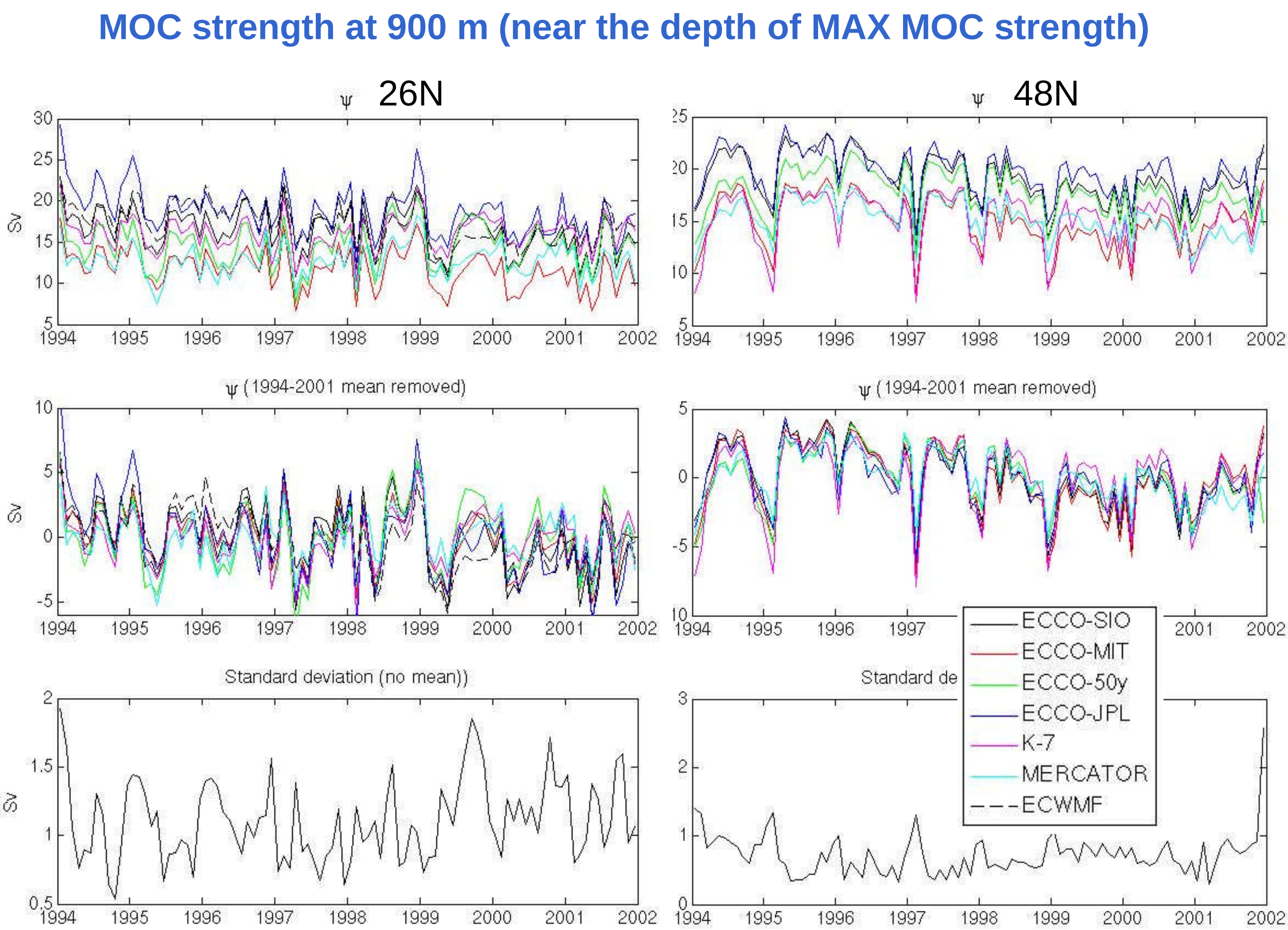

Figure 19: Plot of the AMOC at a depth of 900m at 26°N and 48°N. The mean and seasonal cycle have been removed for each model. Based on results from the GSOP meeting at ECMWF, August 2006. From Lee, CLIVAR GSOP web site - see earlier footnote for http address.

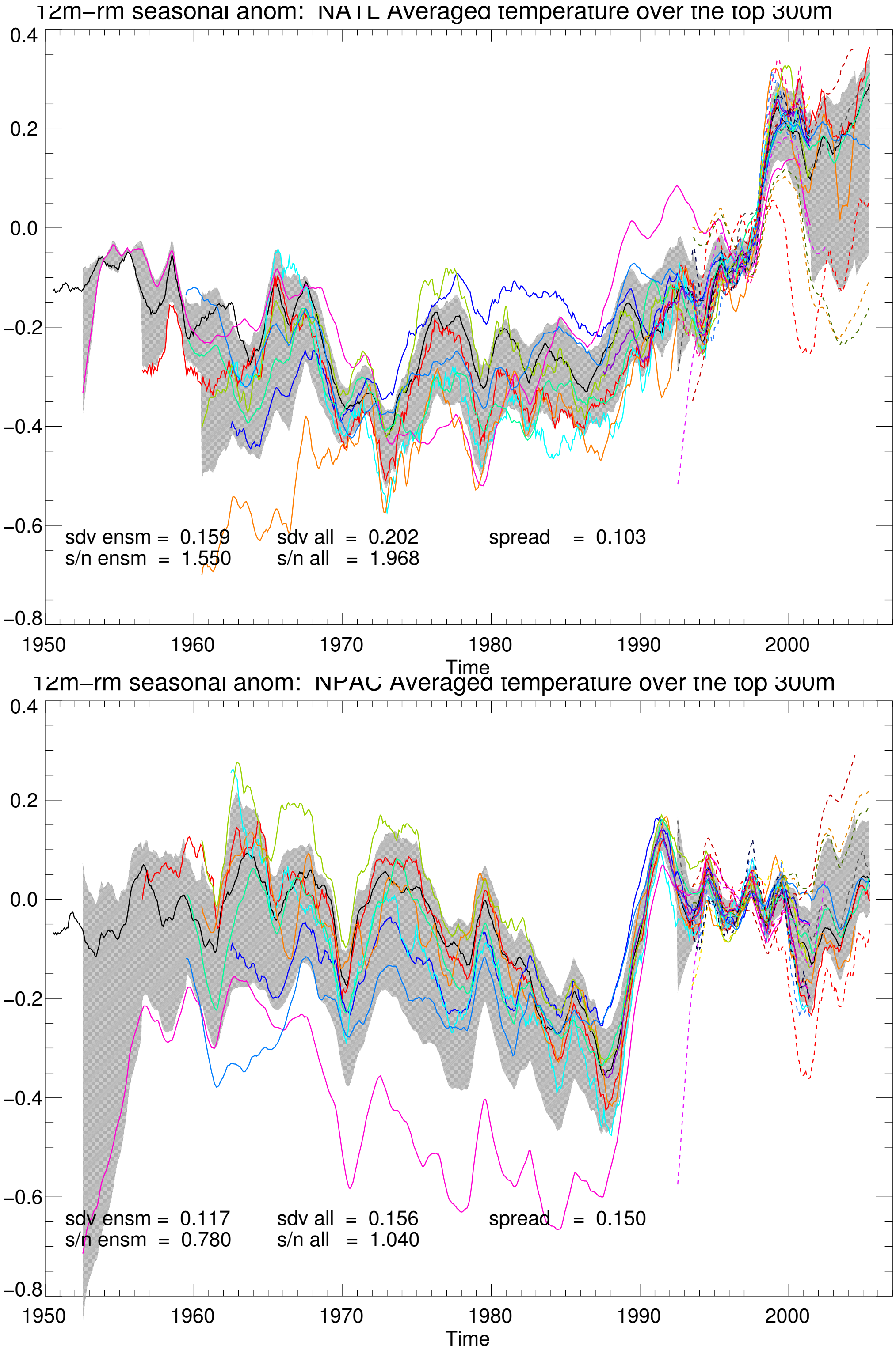

Figure 20: Intercomparison of upper ocean heat content for the midlatitude Atlantic (top) and Pacific (bottom) basins. There is little agreement about the year to year variability although most models indicate an upward trend in the Atlantic and a jump in the Pacific. Based on results from the GSOP meeting at ECMWF, August 2006. From Balmaseda and Weaver, CLIVAR GSOP web site - see earlier footnote for http address.

agreement in the variability. The CERFACS analyses (green) represent 9 different ocean analyses but they all behave so similarly that one can not distinguish that there are 9 members plotted. The lack of variability between ensemble members indicates that the model is far too confident in the analyses. The lack of variability may be related to the fact that there is damping in the analyses. The time scale in low latitudes is that used in ENACT (3 years) as discussed in Davey et al. (2006), but reduces to only 50 days poleward of 60°N, so eliminating variability in the deep sinking regions of the AMOC. There is no known reason for the all three members of the IfM analyses to jump around 1985 but one may speculate that this could be related to the fact that there is no constraint on salinity in that model. The $T-S$ relationship is not applied and consequently unrealistic densities could result, leading to intense changes in deep water formation. The limited ability to initialise ocean models shown by Fig 22 does not instill great confidence in the ability to achieve potential predictive skill.

We now show some forecasts from two of the models viz ECMWF and DePreSys-PP. Extensive results from the other models are not yet available. Given the importance of predicting the Atlantic decadal variability, Fig 23, panels a and b show the AMO index defined in Fig 5, i.e., as the difference between the SSTs in the extra-tropical North and South Atlantic. In this case the re-forecasts show a poorer match to the observed index than with global mean temperature. For this index, DePreSys-PP performs better than ECMWF, especially when the interdecadal variability is considered. Forecasts for the northern component of the AMO index (panels c and d) are worse for ECMWF than for DePreSys-PP, although some hints of predictability for ECMWF can be noticed. Note the large ensemble spread and general overestimation of the variability in DePreSys-PP. The average SSTs over Atlantic ocean north of 10°N (panels e and f) could be considered as another index of the Atlantic decadal variability. While ECMWF performs worse than DePreSys-PP for that index, it succeeds in reproducing the warming trend since the 1980s.

A more in-depth analysis of these integrations should be available in the coming months. The different forecast systems available and the comprehensive set of re-forecasts should make it possible to obtain an estimate of the actual skill of decadal prediction with the current systems.

6 Summary and Conclusions

The large heat capacity of the ocean relative to that of the atmosphere means that it can integrate the white noise forcing of the atmosphere to generate a red spectrum, although this does not necessarily imply long term predictability resulting from natural variability. Predictability studies, however, suggest that many parts of the ocean, notably the North Atlantic, are governed by more predictable processes. Estimates of predictability can be made by running ensembles of hindcasts to assess how quickly they diverge from the true solution. The model is usually perfect, i.e. the same as that used to generate the true solution and the initial conditions are essentially perfect. Other ways to estimate predictability can be made from one or more long integrations by measuring the ppvf. This latter approach has been used by Boer and Lambert (2008) to estimate predictability based on control simulations of IPCC AR4 models. It can also be used on observational records. Although the different approaches and different models give somewhat different estimates, they show considerable similarity in the regions of predictability, all pointing to the North Atlantic. The low frequency variability of the Atlantic, encapsulated in the Atlantic Multidecadal Oscillation (AMO) and the Atlantic Meridional Overturning Circulation (AMOC) have received growing attention over the last few years. Low frequency variability in the Pacific, including the Pacific Decadal Oscillation (PDO) may also indicate predictability but the picture here seems less clear.

Predictability estimates indicate that information on ocean initial conditions influences the forecasts for the order of 5 to 10 years, depending on the predictand. Connections between high latitude sinking in the Greenland

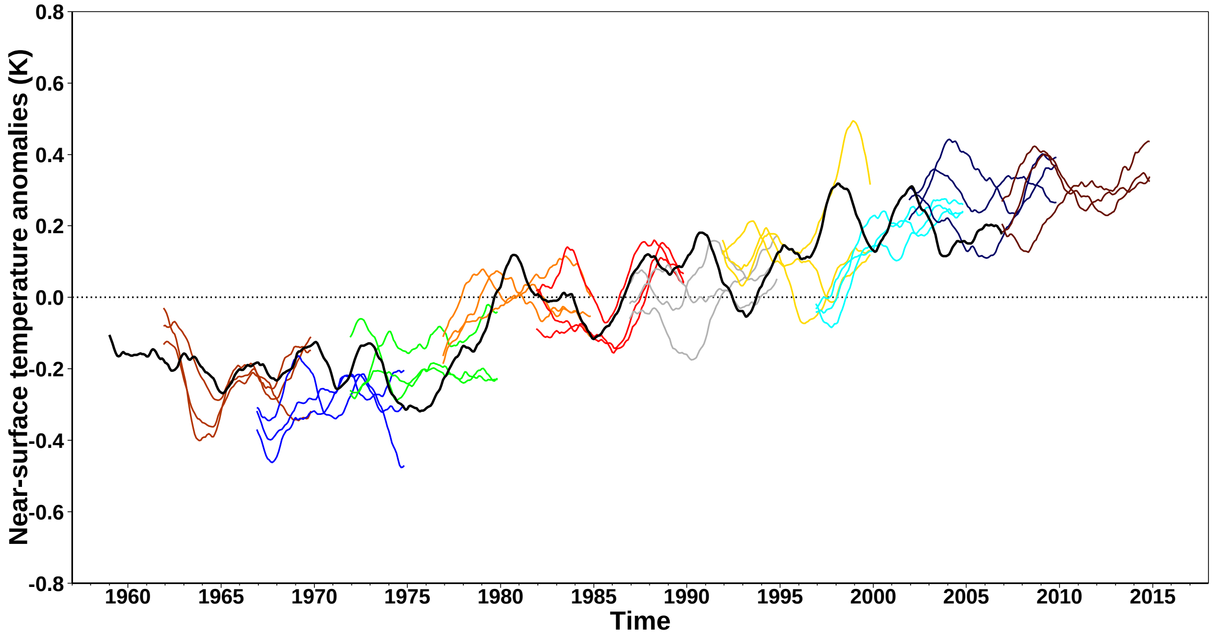

Figure 21: Anomalies of monthly global-mean two-metre temperature from the ERA40 (up to 2002) and operational ECMWF analyses thereafter (black line) and from the three-member ensemble ECMWF (IFS/HOPE) re-forecasts (coloured dashed lines, with a colour for each ensemble start date). The mean systematic error has been removed from the re-forecasts. The annual cycle has been removed and the time series have been smoothed with a two-year running mean.

and Labrador seas, the strength of the AMOC and AMO have been identified and plausible phase lags between them suggested. Connections between the Atlantic and the Pacific have also been suggested, based partly on models and partly on observations. If the AMOC could be predicted, some prediction of European and North American temperature could result. In turn changes in the AMOC may be preceded by changes in the northern Atlantic.

In the last few years, prediction studies have begun. Currently four methods are being tested. Three of these use anomaly initialisation in one form or another. The earliest and most comprehensive is that of Smith et al. (2007). They initiate the ocean component by first analysing it, calculating the anomaly and then adding it to the model ocean after the coupled system has reached a statistically equilibrium state. Likewise for the atmosphere. As a result, model drift during the forecast is reduced since the mean state of the model is balanced. A similar approach was used by Pohlmann et al. (2009) except that the ocean was analysed using a different ocean model to that used to make the forecasts and this leads to some imbalances and inconsistencies. The generation of ensembles is very different. Smith et al. (2007) run a set of ensembles each year whereas Pohlmann et al. (2009) make only one 10-year reforecast each year. In both cases the observed concentrations of radiatively active gases is used. Both highlight the North Atlantic as an area of predictability but the results differ. Radiatively active gases will induce predictability; for example there is a marked trend in temperature as a result of a trend in greenhouse gases. Whereas the ocean initial conditions are important earlier in the forecast, the role of radiative gases will increase the longer the forecast. The timescale over which increases in these gases become dominant depends on the varaiable/process under consideration. For the surface temperature this is only a few years, but considerably longer for the AMOC (Troccoli and Palmer, 2007; Latif et al., 2004).

Both Smith et al. (2007) and Pohlmann et al. (2009) compare the skill of their forecasts with that obtained from unitialised GHG runs. Disappointingly, Smith et al. (2007) find the skill is higher in the uninitialised

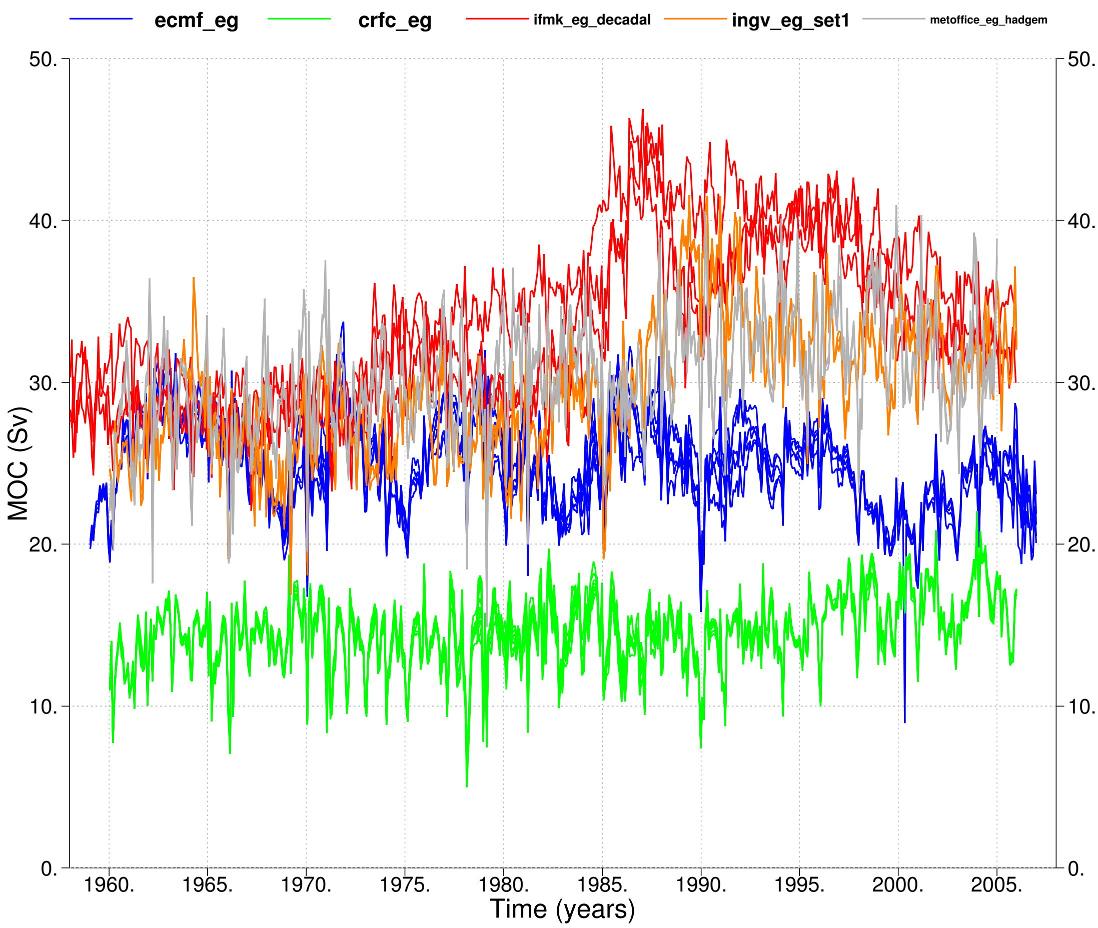

Figure 22: Anomalies of the maximum of the volume transport for 5 ocean model analyses. The Met Office (grey) uses HadGEM2 with the same ocean component as in Smith et al. (2007) and has a three member ensemble of analyses, as does INGV (orange) and IfM (red). ECMWF (blue) has a 5 member ensemble of analyses whereas CERFACS (green) has 9. The transport plotted is a latitudinal average from 10°S to 10°N of the latitude of maximum transport, at the depth level of the maximum. The latitude and depth vary between models.

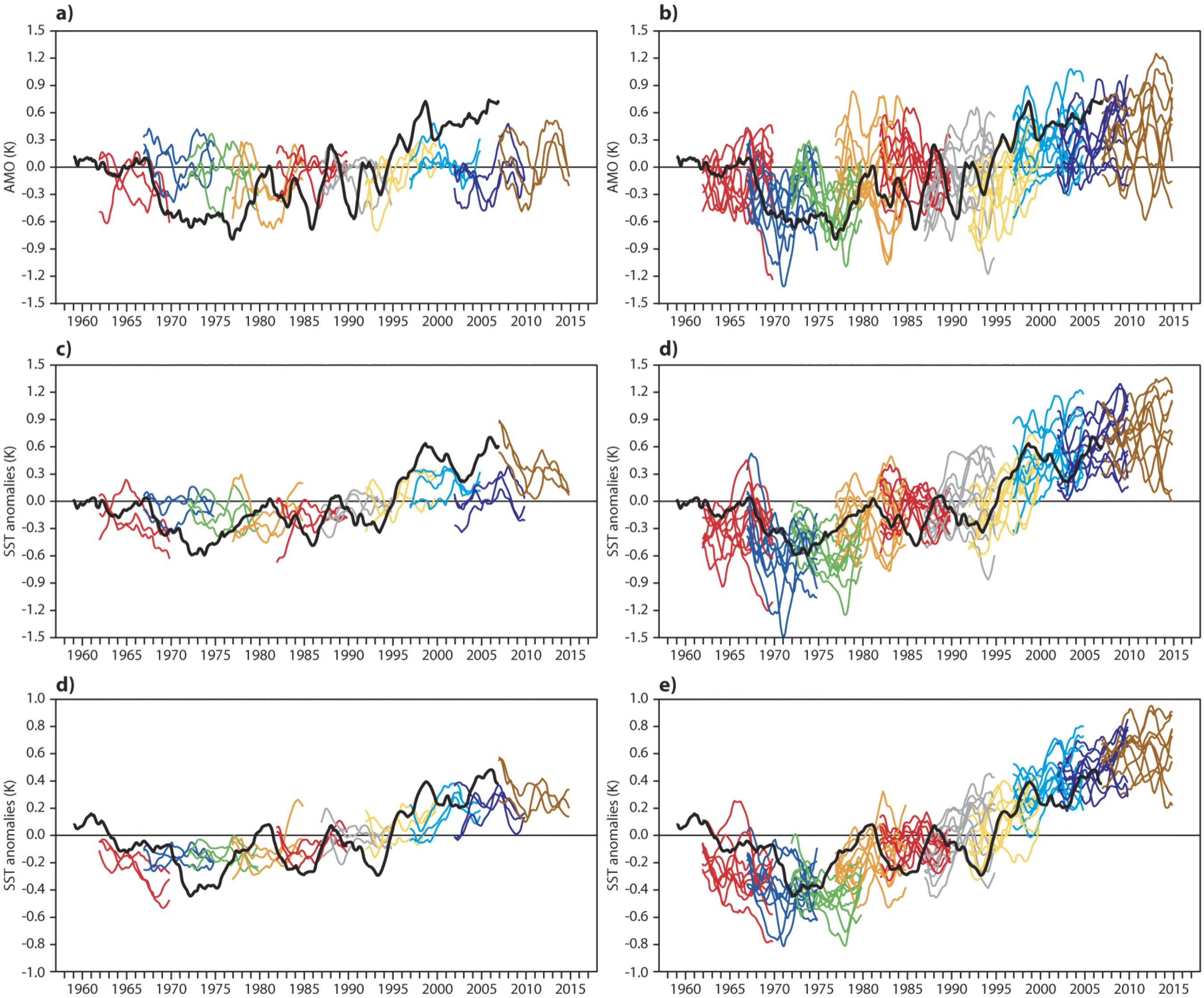

Figure 23: Panels a) and b): AMO index as shown in Fig 5 for the Reynolds SST (black solid line) and the ensemble re-forecasts (coloured dashed lines, with a colour for each ensemble start date). The mean systematic error and the annual cycle have been removed and the time series have been smoothed with a two-year running mean. Panels on the left correspond to ECMWF (IFS/HOPE) and those on the right to DePreSys-PP. Panels c) and d) are similar to a) and b) except for the northern box of the dipole. Panels e) and f) are for the North Atlantic polewards of 10°N.

case than in the initialised case, in the Atlantic sector. However the reverse is true for Pohlmann et al. (2009) and recently, hindcasts made using DePreSys-PP as part of the ENSEMBLES project are more skilful when initialised (Smith, private communication).

A third approach also shows more skill in the initialised case than the GHG runs (Keenlyside et al., 2008). The interesting aspect of this approach is that only SST information is used in order to create the initial state. Even more intriguing is that polewards of 30°, the impact of SST is reduced linearly to zero by 60°. Use of SST in the tropics is a satisfactory approach for seasonal forecasting since the tropical atmosphere responds strongly to SST, though it is inferior to initialising the model with observations of ocean and atmosphere. However, it is surprising that it works to initialise the extratropical atmosphere. There is still uncertainty as to the importance of mid-latitude SST anomalies in generating an atmospheric response. Rodwell et al. (1999) showed that for a certain period they could reproduce some of the low frequency NAO variability in the atmosphere just from SSTs. Not all models respond to extratropical SSTs to the same degree but Rodwell et al. (2004) showed that using several models they were able to capture a remarkably similar response pattern in different models. Additionally, using the HadCM3 model they showed quite a strong response in middle latitudes to tropical SSTs. It is unclear if this is a partial explanation for the Keenlyside et al. (2008) approach.

Like Smith et al. (2007) and Pohlman et al. (2009), Keenlyside et al. (2008) uses an anomaly approach. The only approach not using the anomaly initialisation is that used at ECMWF, where an extension of the seasonal forecast system is employed. The ocean and atmosphere are initialised separately to provide the best analysis of the state of the coupled system, recognising that this state is not balanced and that model initialisation shock and drift might be substantial. As for seasonal forecasting these are removed *a posteriori* using the set of reforecasts. Superficially, the results appear to be of similar quality to those using the anomally approach but no comprehensive quantitative comparison has yet been made. Neither is there a control set of forecasts in which the ocean is not initialised to compare against.

It is clear that for forecasts out to a few years ahead, the initial state of the ocean is important. However, even for the current observation system, there is considerable variation in the estimates of the AMOC strength and variability. Heat content also varies from model analysis to model analysis. For earlier years the uncertainty in ocean analyses is larger. This implies that the skill in decadal forecasting especially based on long records (e.g. from the 60s) will be negatively impacted. Improvements in the observing system may give improved initial conditions but quantifying the impact on forecast skill will be difficult. In the case of seasonal forecasting, model error is probably the biggest impediment to improved forecasting. It is likely that this is a significant impediment to multi-annual forecasting as well.

In Fig 24 Sahel precipitation is plotted showing a strong multi annual cycle over the last sixty years. If this signal could be predicted several years in advance then the potential benefit is very large. The results of Knight et al. (2005, 2006) indicate links between Sahel rainfall and the AMO and AMOC and offer some hope for such prediction.

7 ACRONYMS

AMOC: Overturning Circulation

AMO: Atlantic Multidecadal Oscillation

AMV: Atlantic Multidecadal Variability

ATHC: Atlantic ThermoHaline Circulation

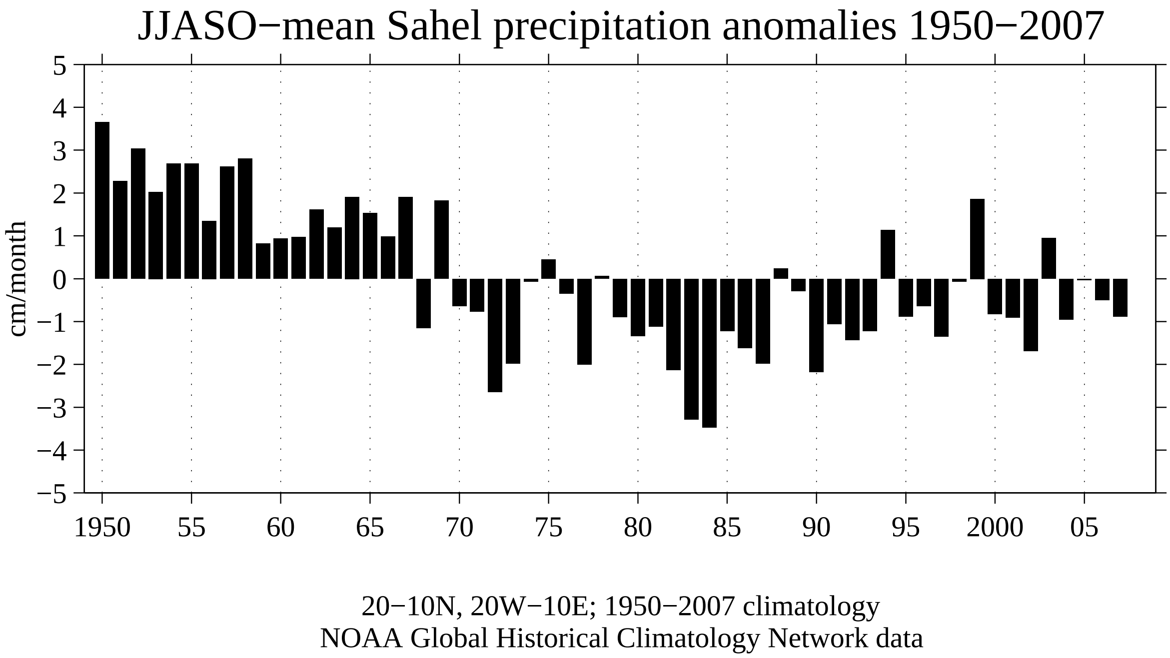

Figure 24: Plot of Sahel rainfall showing low frequency variability. From http://jisao.washington.edu/

ENSO: El Nino Southern Oscillation

GECCO: German ECCO

NAO: North Atlantic Oscillation

NERC: National Environmental Research Council

ORA-S3: The Ocean ReAnalysis System 3, performed at ECMWF

PDO: Pacific Decadal Oscillation

pdf: Probability distribution function

ppvf: Potentially predictable variance fraction

SAT: Surface Air Temperature

SST: Sea Surface Temperature

SSS: Sea Surface Salinity

THC: ThermoHaline Circulation.

TPDV: Tropical Pacific Decadal Variability

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