

Air-Sea Interaction in Seasonal Forecasts: Some Outstanding Issues

Magdalena A. Balmaseda, Laura Ferranti, Franco Molteni

*ECMWF, Shinfield Park, Reading
RG2 9AX, United Kingdom
Magdalena.Balmaseda@ecmwf.int*

ABSTRACT

This work discusses some of the existing deficiencies in the ECMWF seasonal forecasting system related to the misrepresentation of air-sea interaction processes. The discussion focuses on two main topics: the underestimation of peak amplitude of ENSO and the poor skill of seasonal predictions of Northern European summers.

All of the ECMWF seasonal forecasting systems initialized in spring failed to predict the intensification of El Niño 1997 during summer. This failure is attributed to the model inability to sustain the amplitude of an MJO event present in the initial conditions, and failure to propagate it from the Indian Ocean to the Western Pacific. Additional diagnostics suggest that the negative feedback resulting from the model heat flux response to a given SST warm anomaly is too strong. It is argued that this can be a consequence of the model warm bias, which is large enough for the non-linearities to play a role. Experiments with different initialization procedures conducted with the same coupled model indicate that the model warm bias is a direct consequence of the the initial adjustment process.

The low skill of the seasonal forecasting system for the Northern European summers has been investigated. In this area, the anomaly correlation skill score shows negative values, which suggests the presence of a reproducible signal with incorrect spatial phase. The coupled forecasts tend to persist for too long the spring conditions, possibly due to the model inability to reproduce the rapid shallowing of the ocean mixed layer from spring to summer. The lack of prognostic sea-ice model might have contributed to the poor predictions during the last two summers, since the large observed sea-ice anomalies were not accounted for. It is shown that the arctic ice anomalies of the last two summers had a significant effect on the atmospheric circulation over Europe. The model response to the ice forcing is strongly dependent on the background atmospheric circulation, and it is affected by the systematic errors of the coupled model. Experiments demonstrates that the midlatitude sea surface temperature gradient, possibly determined by the representation of the Western Boundary Currents, has a large impact on the atmospheric circulation, and conditions the response of the atmospheric model to the observed ice anomalies.

The most common practice to deal with model error in seasonal forecasts is the a-posteriori removal of the model bias, which assumes that the error in the mean state does not interact with the interannual variability. The results presented here imply that this assumption is inadequate in many cases, and that errors in the mean state of the coupled model are a serious obstacle for further improvements of seasonal forecasts.

1 Introduction

Seasonal forecasting was initiated at ECMWF in 1997, and since then there have been three different seasonal forecasting systems (S1, S2 and S3). All ECMWF seasonal forecasting system is based on a coupled ocean-atmosphere general circulation model that predict simultaneously the lower boundary conditions (namely sea surface temperatures (SST)) and their impact on the atmospheric circulation. In the coupled model, the ocean model provides information about the SST and the ocean current to the atmosphere and wave model, which return fluxes of heat, momentum and fresh water flux. The coupled model is initialized with atmospheric and ocean analyses, and every month is integrated forward in time for 7 months. No relaxation or flux correction is applied during the forecasts. Because of model error, the coupled model drifts towards its own climate with forecast lead time. The drift in SST after 2-4 months can reach 1 degree in the Equatorial Pacific. A common approach is to remove the drift a posteriori: a set of historical hindcasts is performed to provide an estimate

of the model climatological PDF, which is then used for a-posteriori calibration of the forecast results. The a-posteriori bias correction implicitly assumes that errors in the mean will not affect the representation of the interannual variability. A full description of the latest seasonal forecasting system S3 is given in Anderson *et al* 2007 and Molteni *et al* 2007.

The ocean initial conditions are an essential part of the forecasting system, since the predictability at seasonal time scales largely resides in the initial state of the ocean. The ocean initial conditions are produced by assimilating of ocean data into the ocean model forced by analyzed fluxes. The initialization of the ocean and atmosphere is done separately. Although this is a common and practical solution, the separate initialization could produce undesirable initialization shock, which could lead to degradation of the forecasts.

Figure 1 summarizes the main features of the three successive ECMWF operational systems regarding ENSO forecasts. The skill in predicting ENSO has been improving at a steady pace, as can be seen if fig 1a, which shows the root mean square error (rmse) of the forecasts of SST in region Nino3 as a function of lead time. In the last 10 years there has been an improvement in skill equivalent to a gain of 1 month lead-time. The drift and amplitude of the interannual variability (model versus observations) appear in figure 1b and c respectively. S1 had a strong cold bias, consequence of the strong equatorial trade winds. This resulted in a too strong upwelling regime and overestimation of the interannual variability. The introduction of S2 reversed these errors. The equatorial easterlies were on the weak side, resulting in a warm bias. The amplitude of the interannual variability was underestimated. In S3 the amplitude of the interannual variability is improved, although it is still underestimated respect the observations. However, the warm bias is even stronger than in S2, in spite of reduced errors in the mean equatorial wind stress (not shown). The warm bias in S3 is likely resulting from the unbalanced initialization, as will be discussed later.

Despite the steady improvements on the seasonal forecast skill over the years, there are persistent deficiencies which hinder further progress. This work discusses some of the problems that are related to the representation of the air-sea interaction. Section 2 discusses the underestimation of ENSO variability, related to incorrect representation of intraseasonal variability, heat flux response to a given SST anomaly and non linear interaction with the model drift. Section 3 discusses several factors that contribute to the poor skill over Northern Europe, namely the representation of the ocean mixed layer, the lack of prognostic sea-ice and the poor representation of the western boundary currents. The cases presented in this work also illustrate the limits of the linear assumption made in the a-posteriori bias correction.

2 Underestimation of ENSO variability

The 1997-1998 El-Niño was one of the strongest on record. Its onset was predicted by several numerical models, though none fully captured its intensity. This was the case for the ECMWF seasonal forecasting system 1 (S1) which underestimated the intensification during the period June-July 1997 by more than 1K. The subsequent operational systems (S2 and S3) showed similar behaviour: figure 2 shows that forecast for NINO3 SST initialized in April 1997. None of the forecasting systems reproduced the intensification of the anomalies during summer 1997.

Several strong westerly wind bursts (WWB) developed during the onset of the 1997-1998 El-Niño. Figure 3 shows the equatorial evolution of analysed anomalies of SST, sea level, and wind stress for May, June and July 1997. (The SST, wind stress and heat flux data comes from the ECMWF atmospheric analysis system. The sea level is taken from an ocean analysis in ocean data have been assimilated. The anomalies are computed relative to the corresponding 1991-6 analyzed climatology.) Strong WWB, associated to the active phase of the Madden Julian Oscillation (MJO), developed from May to mid-June 1997 and propagated eastward from the Indian Ocean to as far east as 150W, reaching peak values of $0.2 N/m^2$. Although much attention has been devoted to the February-March wind events (for example Legaigne *et al* 2003), the wind event in May-June was also very large. It originated as part of a MJO event in the far western Indian Ocean in April, traversed the Indian Ocean in May, weakened when it reached the Pacific but re-intensified in June. Indeed part of this

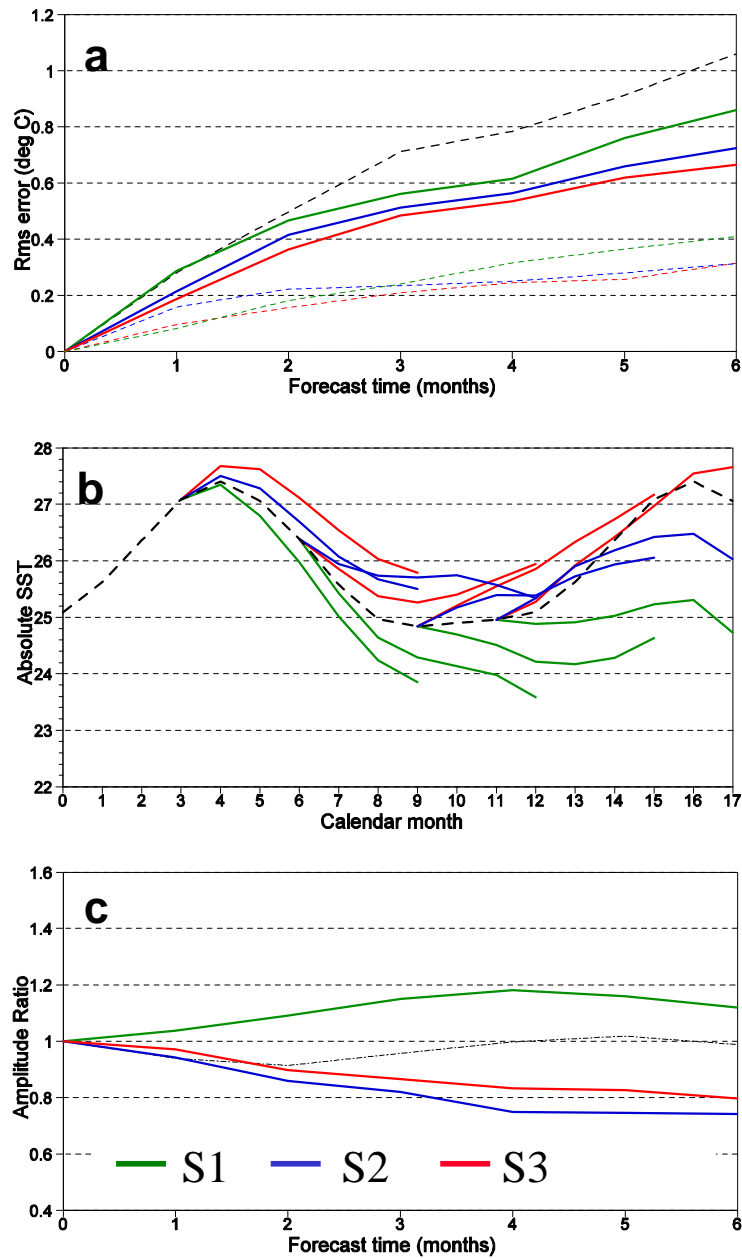


Figure 1: Performance of the successive operational seasonal forecasting systems at ECMWF in terms of SST in region Niño 3. Shown is the SST drift (upper panel), amplitude of interannual variability ratio (model versus observed, in middle panel) and anomaly correlation skill as a function of lead time. The improvement in skill over the last decade is equivalent to a gain of 1 month forecast lead.

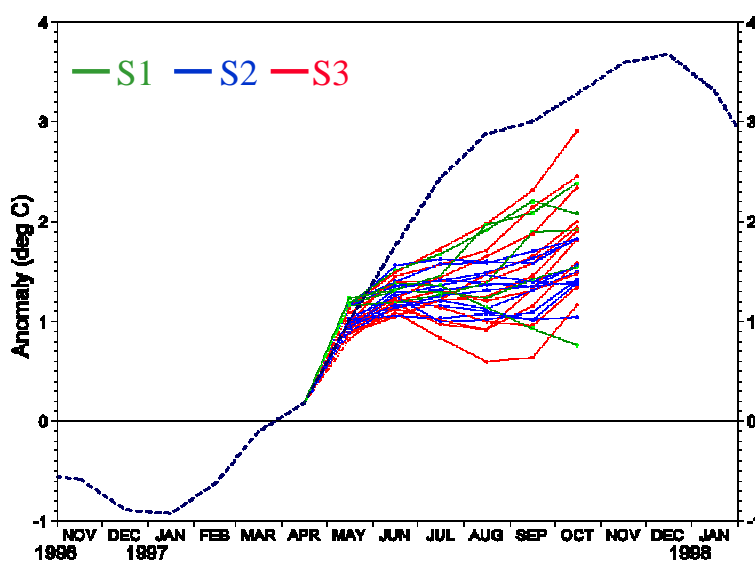


Figure 2: Ensemble of forecasts of SST in Niño3 initialized in April 1997, from the successive operational seasonal forecasting systems at ECMWF. All the forecasts initialized in spring failed to predict the intensification of the anomalies.

westerly wind event passed far into the eastern Pacific. The impact on the ocean state was considerable. The sea level evolution shows eastward propagation of a Kelvin wave generated by the intense westerly wind anomaly in the west Pacific (around 160E- 170E) at the beginning of June. The arrival of this wave in the eastern Pacific coincides with the intensification of the SST anomaly.

In contrast, the seasonal forecasts failed to produce the intensification of SST. Figure 4 shows the equatorial evolution of the S3 forecast of wind stress, sea level and SST from forecasts initialized in May. The figure shows the ensemble mean of anomalies, computed respect the model climatology for the period 1991-1996. The MJO signature is strong in the initial conditions, but the model fails to reproduce its eastward propagation into the Western Pacific and can not sustain its amplitude. After a few weeks the MJO disappears from the model forecasts. The corresponding sea level anomaly associated to the downwelling Kelvin wave is absent, and there is not intensification of the SST in the eastern Pacific during June-July.

2.1 Impact of the MJO and associated WWB

Vitart *et al* 2003, using a prototype of a previous seasonal forecasting system (S2) investigated the impact of the May WWB in the seasonal forecasts. The behaviour of S2 was similar to that shown in fig. 4 for S3. An ensemble of coupled ocean-atmosphere integrations were conducted where the observed wind stress anomalies over the tropical Pacific were added to the wind stress from the atmospheric model. The response to this WWB perturbation appear in fig. 5. Shown are the differences between the perturbed and original forecasts in terms of wind stress (upper left), sea level (lower left), SST (lower middle) and heat flux (lower right). The WWB perturbation produces a Kelvin wave which manifests on a sea level anomaly of 0.15 m, and an SST anomaly of 1.5K. The perturbed forecasts produce significantly better forecasts of SST over the NINO3 region (upper right panel), and the warming in this region exceeds the SSTs in the control run by more than 0.5K. However, the perturbed run still under-predicts the full magnitude of the observed SST anomaly. The WWB only explains about 50% of the observed SST anomaly. The coupled model response to this warming is characterized by a negative feedback in terms of heat flux exceeding values of 120 W/m^2 , that acts to reduce the SST anomaly.

The coupled model showed two main deficiencies regarding the representation of the MJO and the associated WWB: the failure to sustain its amplitude and the inability to propagate the event from the Indian Ocean into the Western Pacific. Vitart *et al* 2007 show that the amplitude of the intraseasonal variability in the coupled model is largely dependent on the parameterization of convection. Regarding the propagation of the MJO, Woolnough *et al* 2007 have shown the ocean mixed layer is instrumental in the propagation of the MJO. In their study they conclude that the vertical resolution of the ocean mixed layer should be fine enough as to allow the representation of the diurnal cycle and the fast response of SST to the intense convective cooling. In the case study presented here, the MJO responsible for the intensification of the SST was already present in the initial conditions, and the model failed to propagate it eastward. It is therefore likely that any improvements in the representation of the ocean mixed layer that contribute to the better propagation of the MJO events will translate in better seasonal forecast skill.

2.2 Impact of the warm bias

The results discussed above indicate that the absence of westerly wind variability only partially explains why the ECMWF seasonal forecasting system failed to predict the strong warming in the NINO3 region when starting on 1st May 1997. Even when the westerly wind events are included, the NINO3 forecasts are still well below observations. This is likely the consequence of the strong dumping effect of the heat flux response. The ratio between heat flux anomalies and SST anomalies in figure 5 is about $80 \text{ W/m}^2/\text{K}$, much higher than that from the atmospheric analysis (which ranges from 20 to $30 \text{ W/m}^2/\text{K}$). The amplitude of the sea level anomaly is similar to that of the analyzed fields, suggesting that the dynamical response of the coupled model to the wind anomaly is correct.

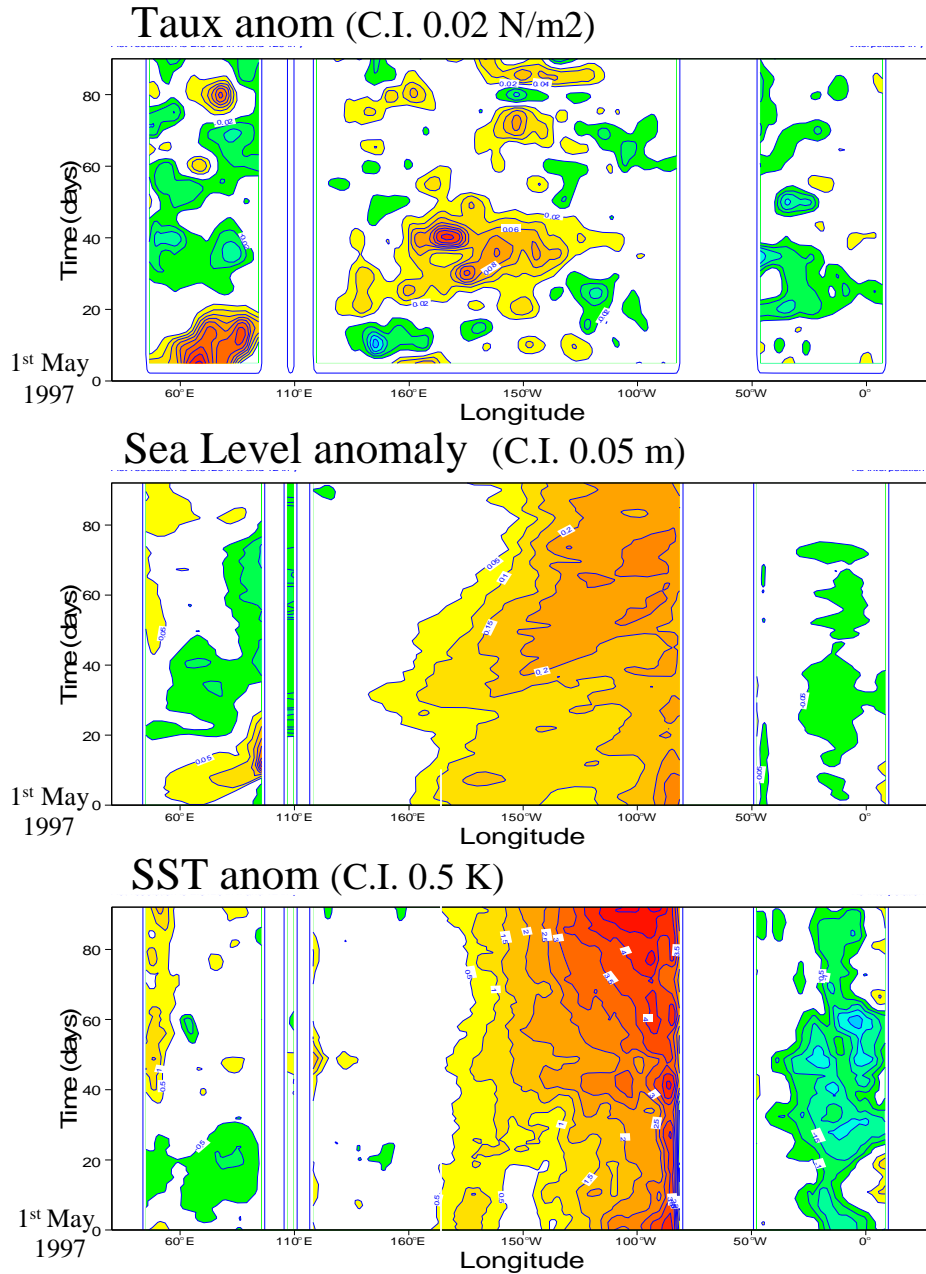


Figure 3: Time evolution (from 1st May 1997 to 1st August 1997) of the analysed anomalies of equatorial a) wind stress, b) sea level, and c) SST; all are with respect to the analyzed 1991-1996 climatology. C.I. is 0.02 N/m², 0.05 m, and 0.5 K respectively.

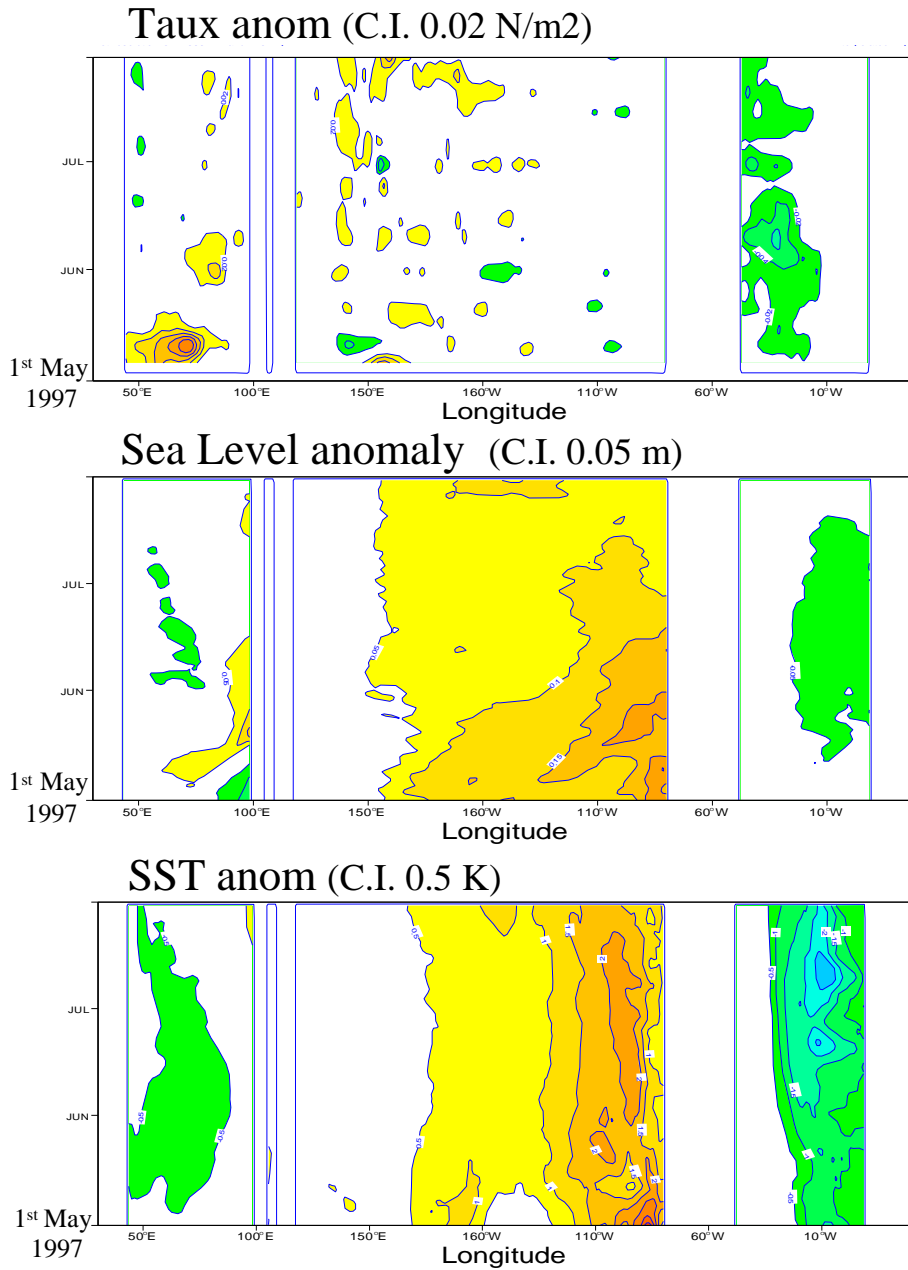


Figure 4: Time evolution (from 1st May 1997 to 1st August 1997) of the forecast anomalies of equatorial a) wind stress, b) sea level, and c) SST; all are with respect to the analyzed 1991-1996 climatology. The anomalies are the 5-member ensemble mean of forecasts from the S3 operational system. C.I. is 0.02 N/m², 0.05 m, and 0.5 K respectively.

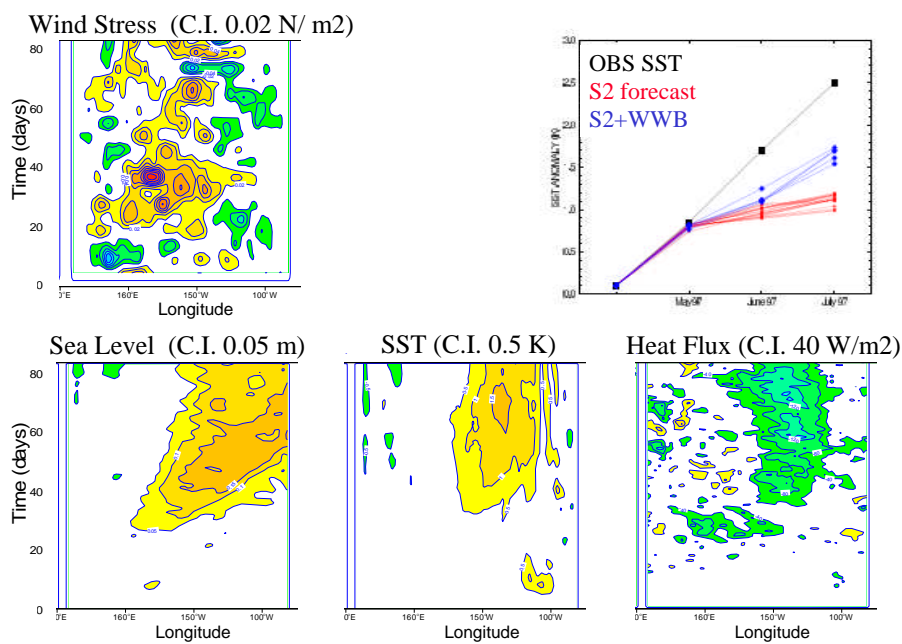


Figure 5: Time evolution of the response of the coupled model to the perturbation in wind stress shown in the upper left panel. The lower panels show the ensemble mean difference in equatorial Pacific between the perturbed and unperturbed forecasts in sea level (lower left), SST (lower middle) and heat flux (lower right). Start date is 1st May 1997. The upper right panel shows the forecast of SST from the unperturbed (red) and perturbed (blue) forecasts.

The previous experiments were performed with an earlier version of the coupled model. In the coupled model used in the current seasonal forecasting system (S3), the heat flux dumping is about $40 \text{ W/m}^2/\text{K}$, which, although still higher than the observed relation, is not as strong as in the earlier version. This could explain why the SST interannual variability is higher in S3 than in S2 (fig 1).

Balmaseda and Anderson 2009 show that that the procedure to initialize the ocean, as well as having a strong impact on the forecast skill, also affects the drift and the amplitude of the interannual variability of the coupled model. Figure 6 shows results from the same coupled model used in S3 with two different initialization strategies, which differ on the amount of observational information used to initialize the coupled system. The red curves are for the procedure used in S3, which uses the most observations available (ocean and atmosphere observations). The green curves are for an experiment where only SST information has been used to initialize the coupled system. Shown are the results for region NINO3, in the Eastern Pacific. The initialization can change the sign of the drift (fig 6a) and the amplitude of interannual variability (fig 6b), which is higher for colder bias. The warm bias in the S3 initialization is caused by quick dynamical adjustment in the form of a downwelling Kelvin wave, since there is an imbalance between the slope of the thermocline in the initial conditions and the strength of the winds in the coupled model. In spite of the warm bias, Balmaseda and Anderson show that using more observational information improves the forecast skill. Forecast improvement is observed in most of the individual forecasts, but not for all. A remarkable exception are the forecasts for the peak of El Niño 1997/8. Figure fig 6c shows an example of forecasts initialized in October 1997 (close to the peak of ENSO). The initialization does not influence much the forecasts initialized in May 1997 (not shown), which are characterized by the absence of the WWB as discussed earlier. However, the forecasts initialized in October 1997 are very sensitive to the initialization. The procedure used for S3 is unable to sustain the peak amplitude of ENSO. However, the initialization using only SST is able to amplify the ENSO conditions even further. The heat flux dumping is slightly weaker in cold bias case (about $20 \text{ W/m}^2/\text{K}$). One possible explanation for the different behaviour is the effect of the mean drift: the heat flux dumping is stronger in the forecasts with the warmer bias, which would imply the existence of non-linear interactions between the mean state and the variability. Although more work is needed to understand the specific nature of this non-linearity, it is clear that the initialization procedure has a large impact in the coupled model drift and amplitude of interannual variability. Better initialization methods are needed to avoid unwanted drifts caused by initial adjustment, while still making use of all available information while still producing balance ocean-atmosphere initial states.

3 Seasonal Forecasts of European Summers

Figure 7a shows the skill of the S3 seasonal forecasts of JJA Z500 from May starts. The skill is measured in terms of the anomaly correlation, based on an ensemble of forecasts covering the period 1981-2005. For other skill scores see http://www.ecmwf.int/products/forecasts/d/charts/seasonal/forecast/seasonal_range_forecast/. The significant negative values of the anomaly correlation over Northern Europe are tantalizing, since they would imply that some degree of predictability is possible if the spatial phase of the signal is reproduced. The negative correlation indicates that the coupled model is unable to reproduce the correct spatial phase of the signal. This section investigates some of the possible reasons for this failure.

3.1 Ocean mixed layer

Figures 8, from van Oldenborgh 2007 show the anomaly correlation between the anomalies of 2m temperature (T2m) in spring and the following summer, as a measure of the persistence of the spring anomalous condition. The figure shows results for ERA-40 and for the multi-model from the DEMETER integrations. Over the north subtropical Atlantic and Europe, the models show stronger persistence than the reanalysis data.

A possible reason for the extended persistence of spring conditions into the summer reside in the ocean mixed layer. The observed mixed layer in the subtropics shallows very rapidly during the spring, so that by June has

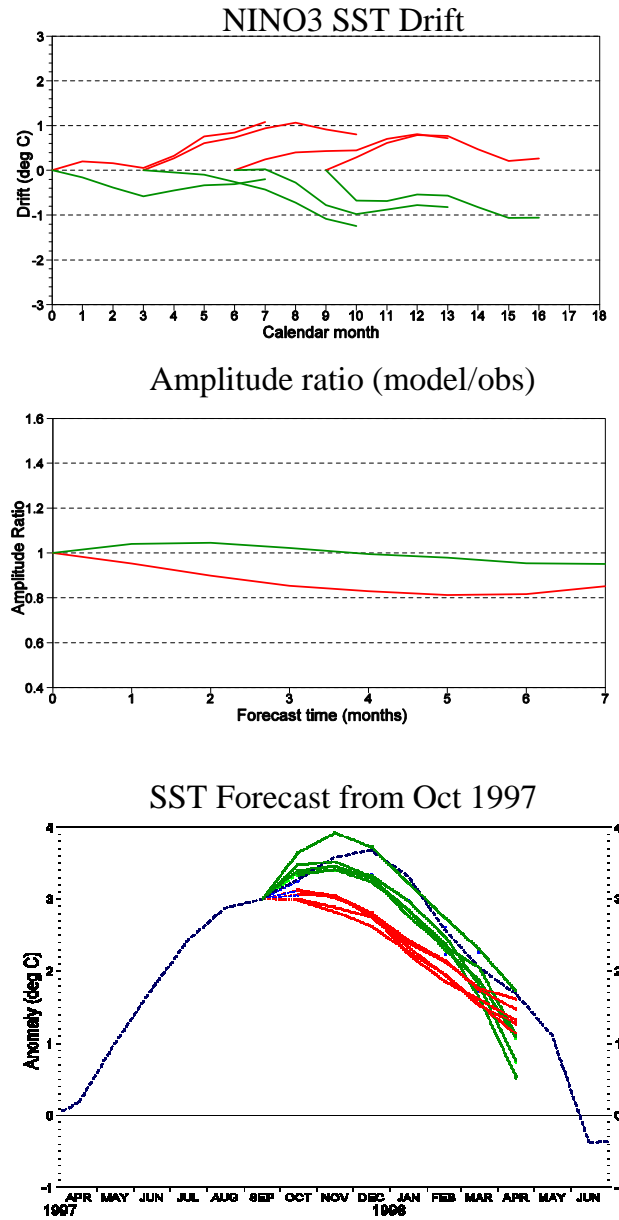


Figure 6: Impact of initialization procedure on the SST drift (upper panel) and amplitude of interannual variability (middle panel). Results are for the region NINO3. The lower panel shows the forecasts of SST in NINO3 from October 1997. The green lines are for the initialization procedure that only uses SST. The red lines are for the initialization procedure used by the ECMWF S3.

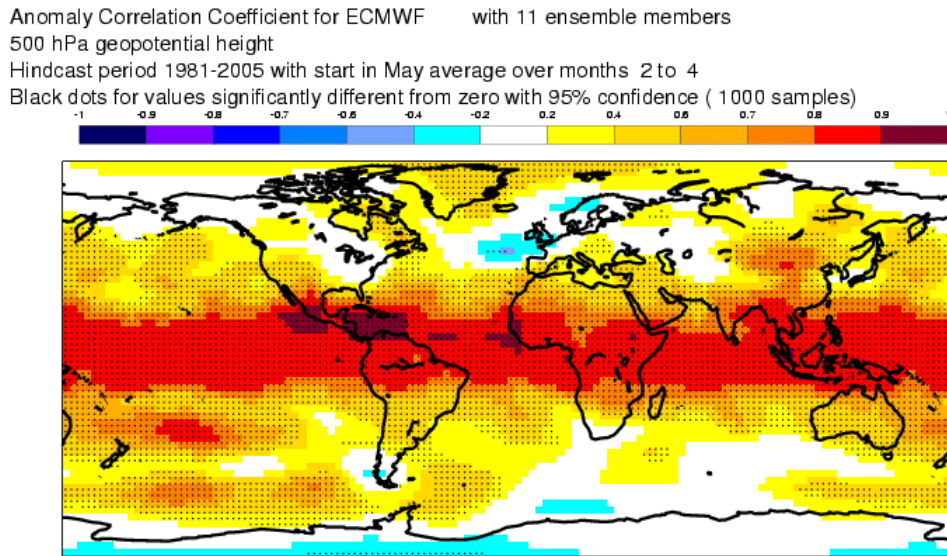


Figure 7: Anomaly correlation for predictions of Z500 anomalies for JJA from May starts from the ECMWF S3 seasonal forecasting system, from http://www.ecmwf.int/products/forecasts/d/charts/seasonal/forecast/seasonal_range_forecasts/. The anomaly correlation has been estimated by an ensemble of forecasts for the period 1981-2005. The areas where the correlation is significantly different from zero at the 95% level are shaded.

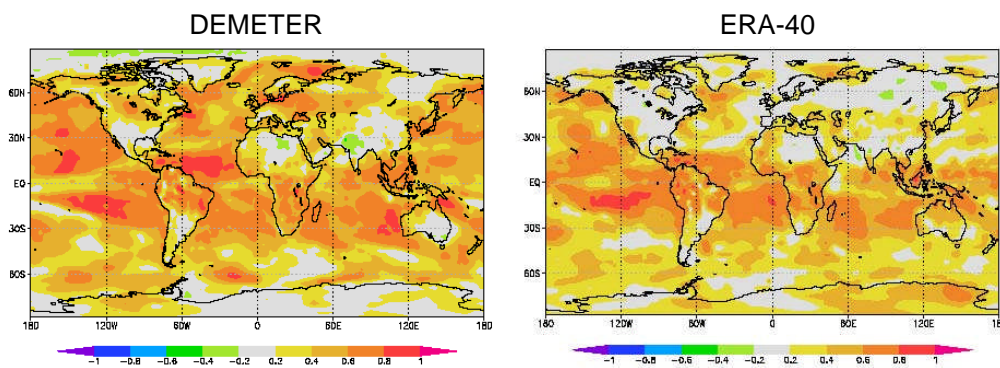


Figure 8: Persistence of T2m spring conditions in ERA 40 (left) and in the DEMETER integration (right). The persistence is measured as the correlation between the spring and summer anomalies. From G.J. van Oldenborgh 2007

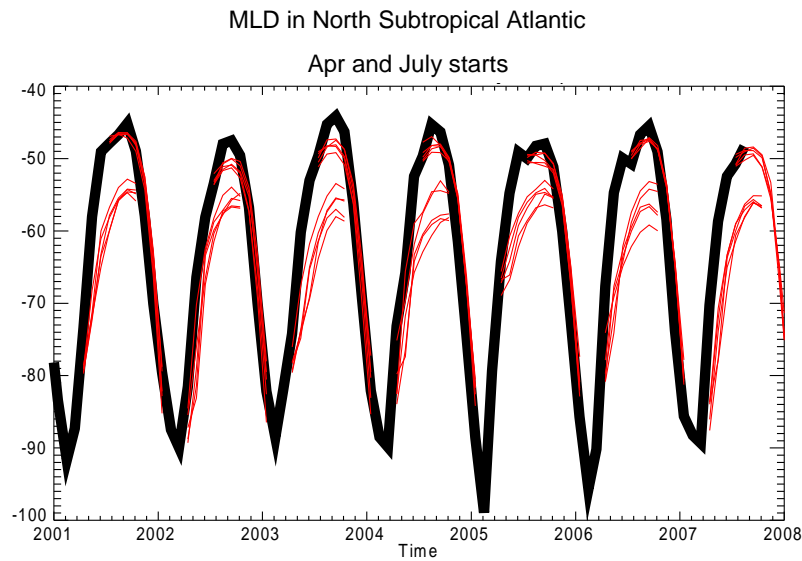


Figure 9: Time series of the mixed layer depth in the North Subtropical Atlantic (5N-28N), from the ocean analysis (black) and from seasonal forecasts initialized in April and June. The forecasts initialize in April systematically overestimate the depth of the mixed layer depth

reached minimum thickness (40 to 50 m). This thin mixed layer has little thermal inertia and reacts quite rapidly to the surface forcing. The rapid shallowing of the mixed layer could explain the little persistence from spring to summer in observations. In contrast, the coupled model fails to reproduce the rapid mixed layer shallowing. Figure 9 shows the seasonal prediction of mixed layer depth (MLD) in the north subtropical Atlantic from April and June initial conditions (red curves). The forecasts starting from April systematically overestimate the MLD (compared with the analysis in black), which could explain why in the coupled model the spring conditions persist for too long. The forecasts initialized in June, when the mixed layer is already shallow, are not biased.

The reasons for the model failure to reproduce the rapid shallowing of the ocean mixed layer need to be investigated further. They could be due to deficient parameterization of mixed layer processes in the ocean model, poor vertical resolution, lack of ocean colour... The problems may still be of atmospheric origin, such as inadequate atmospheric fluxes. However, experiments with a finer resolution ocean mixed layer model (Vitart, this volume) show that the representation of the ocean mixed layer has some impact on the forecast of temperature over the Atlantic region for forecasts initialized in May. Further experimentation is needed to establish the impact of the mixed layer on the forecast skill.

3.2 Effect of recent Arctic ice anomalies

In S3 there is not a dynamical ice model. Instead, the ice concentration is persisted during the first 15 days, after which damped persistence is used for an additional 15 days, when climatological values are used. The lack of a prognostic ice model may be a shortcoming of the current forecasting system. The summers of 2007 and 2008 have seen unprecedented reduction in the Arctic sea-ice extension (figure 10), which might have been responsible for the anomalous atmospheric circulation over the Euro-Atlantic Sector in the last 2 years.

Figure 11 shows the observed Z500 anomalies during July-August of 2007 and 2008. The anomalies have

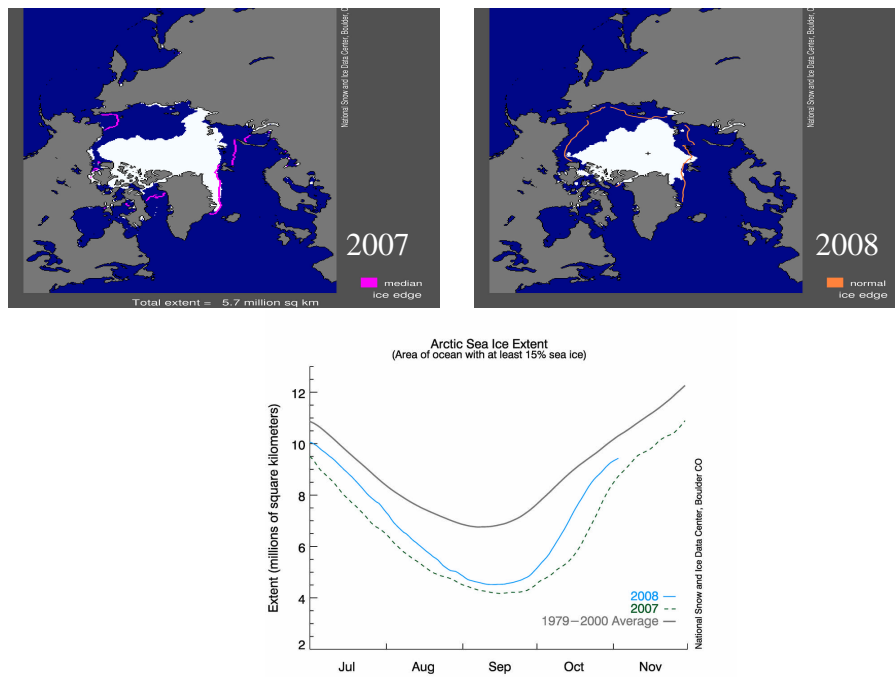


Figure 10: Arctic sea ice extension for the summer 2007 (upper left) and 2008 (upper right). The lower panel compares the time evolution of the ice extension for the last two years with that of the long term climatology. Figures are from the National Snow and Ice Data Center.

been computed respect the 1979-2001 ERA40 climatology. It has been argued that the sea-ice interannual variations are, to a large extent, a response to the atmospheric forcing (Haas and Eicken 2001), especially to the wind forcing. Slingo and Sutton (2007) have speculated that the anomalous atmospheric circulation over the Arctic in the summer of 2007 might have been a consequence of La Niña conditions. L'Heureaux *et al.* 2008 argue that the large ice anomalies of 2007 were a consequence of the strong anticyclonic circulation over the Arctic, dismissing the connection with tropical forcing arising from La Niña conditions. The anomalous atmospheric exhibits consistent Arctic high during 2007 and 2008. The negative centers of action over North-Western Europe and North-Eastern America are also present in both years. The anomalous Z500 pattern, and its associated imprint on SLP (not shown) is consistent the the response pattern of the atmospheric circulation to the long-term trends in arctic sea-ice found by Gerdes *et al.* 2006., Alexander *et al.* 2004; Deser *et al.* 2004, Deser and Teng 2008. Although all these works refer mainly to winter conditions, their results would suggest that the atmospheric circulation observed during the summers of 2007 and 2008 could be a consequence of the ice anomalies, rather than a forcing agent.

To evaluate the impact of the observed ice anomalies in the Northern Hemisphere (NH) atmospheric circulation, two sets of simulations with an atmospheric General Circulation Model (GCM) were carried out. The first set is forced by daily values of the analyzed ice cover, while the second set uses the same scheme as the S3 seasonal forecasting system, e.i., daily values of climatological ice cover after 15 days into the integrations. Both sets have been forced by the prescribed (observed) values of SST. The experiments, consisting on 40-ensemble members each, were initialized in May 2007 and 2008, and were integrated forward for 5 months forced by the corresponding values of daily SST. Figure 12 shows the difference in arctic sea-ice concentration between the two sets of experiments for 2007 and 2008.

Figure 13 shows the anomalous surface heat forcing resulting form the anomalous ice concentration in 2007, as measured by the difference in surface heat flux between the experiments with observed and climatological ice cover. The heat flux anomaly reaches maximum positive values (heat into the ocean) during July and August, where the melting of the sea-ice leads to the reduction of the albedo and to an excess of solar radiation into

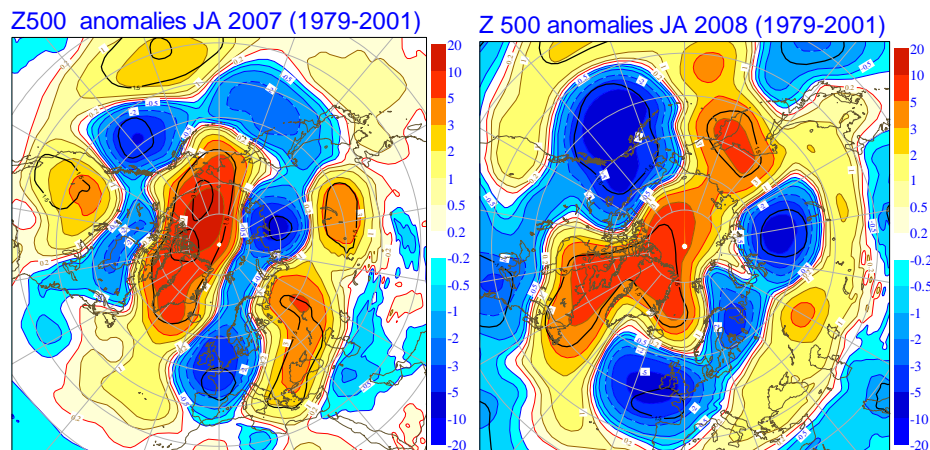


Figure 11: Observed Z500 anomalies during July-August of 2007 (left) and 2008 (right). The anomalies are computed respect the 1979-2001 ERA40 climatology.

the ocean. By September however the heat flux anomaly changes sign, and the contribution of the latent heat becomes dominant, with a net heat flux from the ocean into the atmosphere (negative values). The heat flux that appears over the North Pacific and Atlantic in August and September is a consequence of the atmospheric circulation rather than direct SST forcing. The evolution of the heat fluxes is similar in 2008 (not shown).

The anomalous heat surface forcing during July and August due to the ice reduction has a significant impact in the atmospheric circulation. Figures 14 shows the difference in Z500 between the experiments with observed and climatological ice for 2007 and 2008. Although there are differences between the two years, the response in both cases is quite consistent, characterized by a positive anomaly over the Arctic and a negative anomaly over the North-Western Europe and North-Eastern sector of America. The anomalies are modest in size, but significant. The pattern of the response to the ice anomaly is also consistent with the observed anomalous atmospheric circulation over the Arctic and North-Atlantic sector shown in figure 11.

To assess the impact of the ice anomalies in the coupled model, similar sensitivity experiments were conducted, but this time the SST evolution was predicted by the coupled model, instead of being prescribed from observations. The coupled experiments also consisted on 40 ensemble members, initialized in May 2007 and 2008 and integrated forward for 5 months. Surprisingly, the response to the sea-ice anomalies of the coupled simulations (see later in figure 17) is very different from the response of the uncoupled (or forced) simulations (figure 14), in spite of the surface fluxes associated to the ice anomaly in coupled and uncoupled mode being very similar (not shown). One possible explanation for the different response resides on the non-linear nature of the atmosphere. This idea is explored in the following section.

3.3 Role of Western Boundary Currents

As the coupled model is not perfect, the SST predicted by the coupled model will have errors. The differences between model and observed SST for predictions initialized in May 2008 appear in figure 15, together with the resulting difference in heat flux forcing. The largest differences appear in the regions of the western boundary currents, but they also affect the basinwide structure of the SST. The pattern of error is consistent with the misrepresentation of the Gulf stream in the coupled model: it fails to separate from the coast at cape Hatteras,

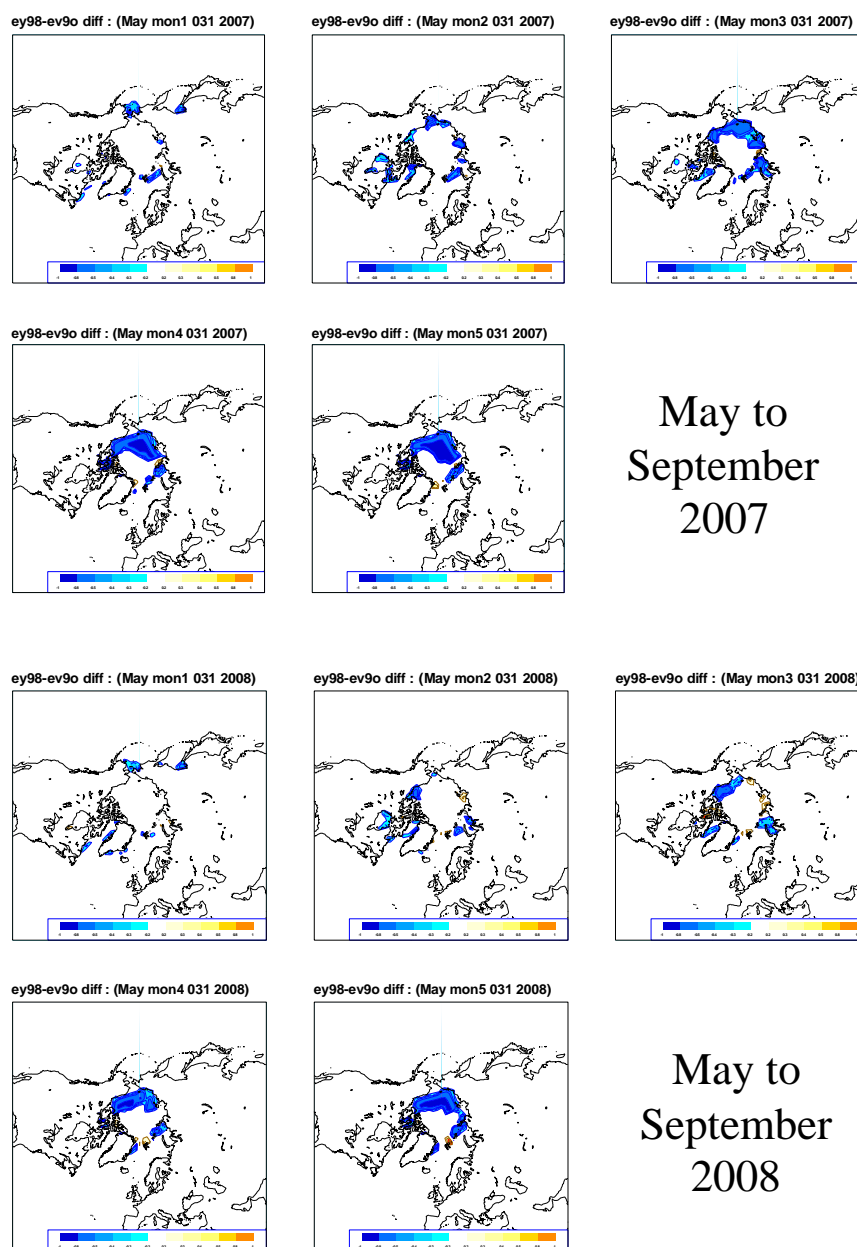


Figure 12: Difference in sea ice concentration between the experiments with prescribed and climatological ice during May-September of 2007 (upper figures) and 2008 (lower figures)

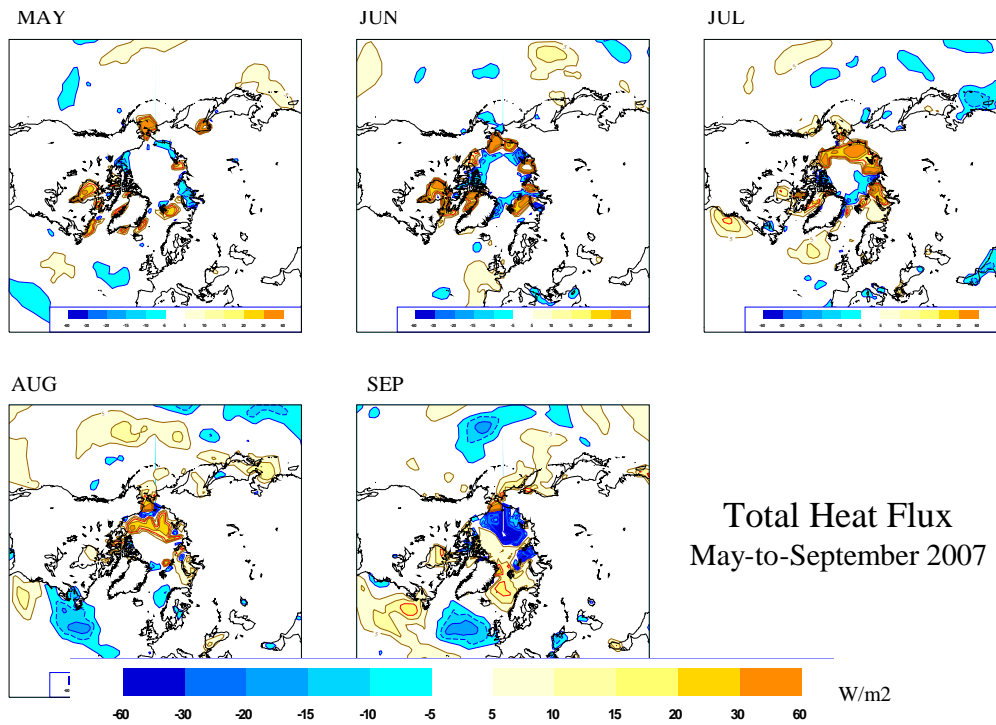


Figure 13: Differences in the total surface heat flux between the experiments with prescribed and climatological ice during May-September 2007.

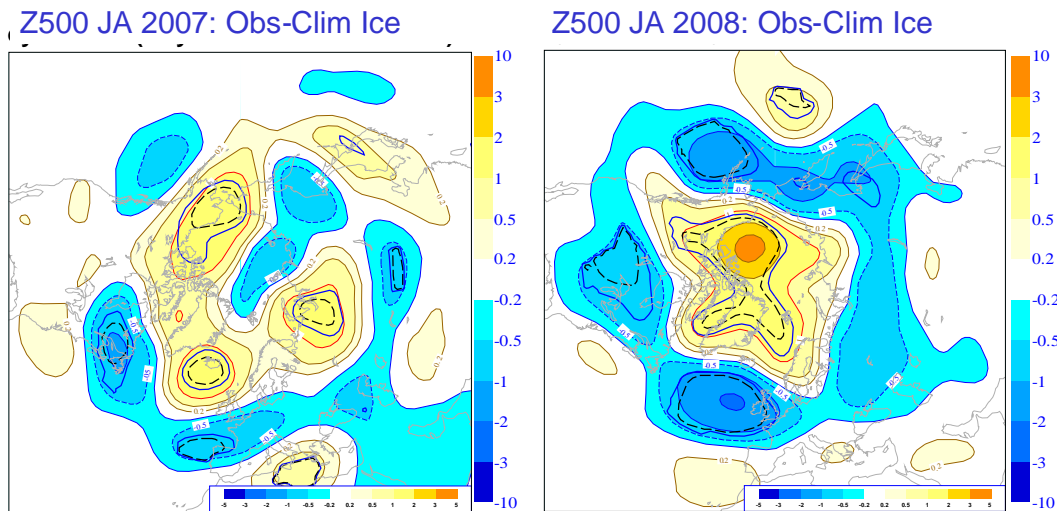


Figure 14: Impact of the summer ice anomalies of 2007 (left) and 2008 (right) on the July-August atmospheric circulation, as measured by the ensemble mean difference in Z500 between two experiments in which the atmosphere model is forced by the analyzed ice coverage and by climatological ice respectively. The experiments, with 40 ensemble members each, were initialized in May and run for the 5 months forced by observed SST. Units are dam. The 90% and 95% significance level are shown by the thick blue and dashed-black contours.

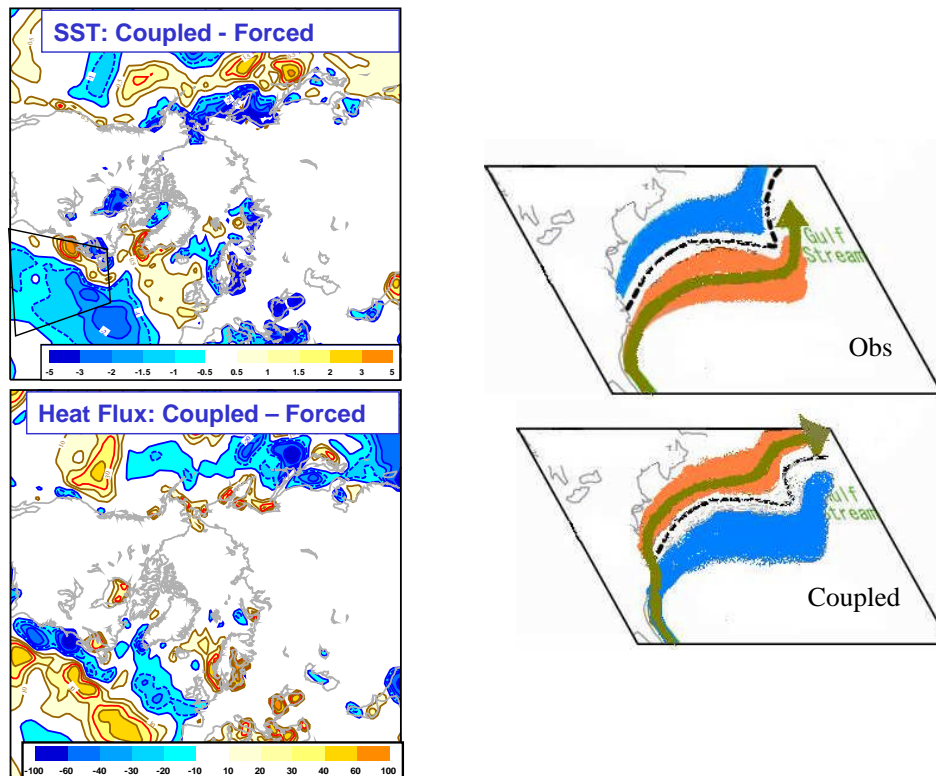


Figure 15: Difference in the SST (upper left) and heat flux forcing (lower left) between the coupled and forced experiments for 2008. The pattern of difference is consistent with errors in the representation of the Gulf Stream as depicted schematically on the right. The square in the SST panel shows the area used for the partial coupling experiment.

and penetrates further north than in observations (see schematic figure in the right panels of figure 15). As a consequence, in the coupled model there is too much heat transported north of along the North-American coast, and there is not enough heat transported meridionally towards the Mid-Atlantic Ridge. The heat flux difference associated to the displacement of the Gulf Stream manifests as strong dipole, with too much latent heat flux being released into the atmosphere over the areas of warm SST: near the coast in the coupled experiment and towards the Mid-Atlantic Ridge in the uncoupled case. The misrepresentation of the Gulf Stream, and more generally, of the western boundary currents, is a common error in climate models, which tend to use relatively coarse resolution in the ocean (about 1 degree).

The left panel of fig 16 shows the difference in Z500 between the coupled and forced integrations for 2008. The curvature of the Z500 surface is quite different in coupled and forced mode, and the differences exhibits an annular structure: the coupled model shows higher values over the tropics and a sharper decline at mid latitudes, resulting in lower values of Z500 over the 40N-60N latitude band, and higher values over the Arctic. Experiments conducted for the years 1987-2007 show that the pattern of differences between coupled and uncoupled simulations shown in fig 16 is robust, and is also linked to the systematic drift of the coupled system.

The large heat flux exchange in the North Atlantic seen in figure 15b is likely to affect the atmospheric circulation. To find out how the western boundary affects the atmospheric circulation, an additional experiment with partial coupling was conducted, where the observed SST were prescribed only over the Gulf Stream area (30N-60N, 80W-30W), shown by the polygon superimposed in left panel of figure 15. Everywhere else, the model is fully coupled. The partial coupled integrations were initialized in May 2008, and consist on 20 ensemble members. The effect of the Gulf Stream in the atmospheric circulation, measured as the differences between

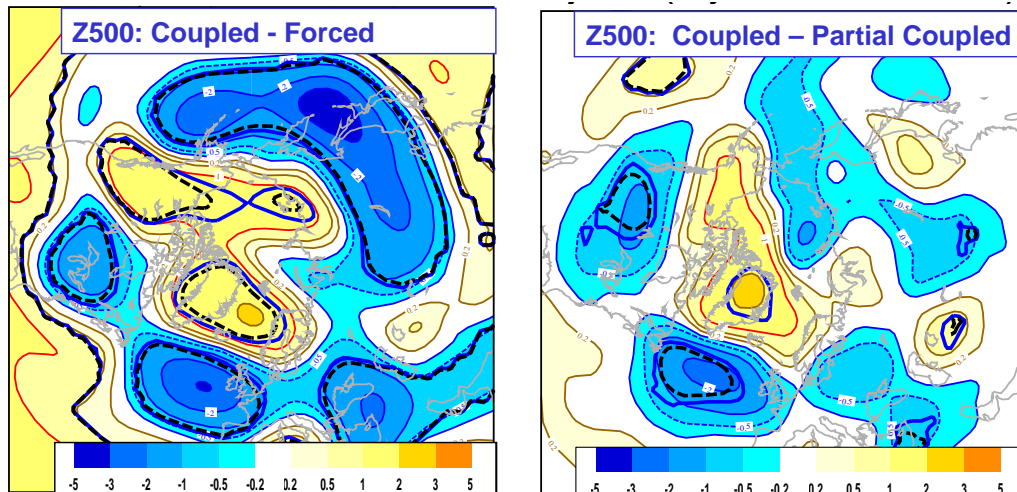


Figure 16: Difference in the atmospheric circulation between uncoupled and coupled mode, in terms of Z500 (left). The right panel shows the impact in Z500 of correcting the Gulf stream, as the difference between the experiment with partial coupling and the coupled model. Units are dam. The 90% and 95% significance level are shown by the thick blue and dashed-black contours.

the ensemble mean of coupled and partial-coupled experiment, is shown in the right panel of figure 16. By correcting the SST over the Gulf Stream area it is possible to account for most of the differences between coupled and forced integrations over the Euro-Atlantic sector and Greenland area. The impact is not so significant in the remote areas of Siberia and North Pacific.

The response of the coupled model to the ice anomaly for 2008 is shown in the left panel of figure 17. The response is very different from that of the forced model (right panel of figure 14), being almost out of phase over the Arctic and Euro-Atlantic sector. If the response to a given ice anomaly is flow-dependent, the different mean state in the coupled and forced mode will lead to different response to the anomalous ice forcing. This hypothesis is tested by investigating the effect of the ice anomaly in the partial-coupling experiment. The sensitivity to the 2008 ice anomaly in the partial-coupling experiment appears in the right panel of figure 17. By correcting the values of SST over the Gulf Stream the atmospheric response to the 2008 ice anomaly gets closer to that of the forced model, with high values of Z500 over the Arctic, and a low over North-Western Europe.

The sensitivity of the response to the ice anomaly to the background atmospheric circulation highlights the strong non-linear nature of the atmospheric response to a given forcing. The results indicate that the skill of current seasonal forecasts system over the Euro-Atlantic sector may be limited by the deficient representation of the Gulf Stream. These results can have far more reaching implications: they would imply that accurate seasonal and decadal predictions and climate projections require the representation of the Gulf Stream and other western boundary currents, which is not done correctly by the climate models used in climate predictions.

4 Summary and Conclusions

In spite of the advances on ENSO forecast over the years, the prediction of the peak amplitude of the El Niño 1997/1998 remains elusive. Coupled forecasts started on 1 May 1997 underestimate the amplitude of the NINO3 anomalies in July by more than 1K. One of the reasons for this error is the inability of the coupled

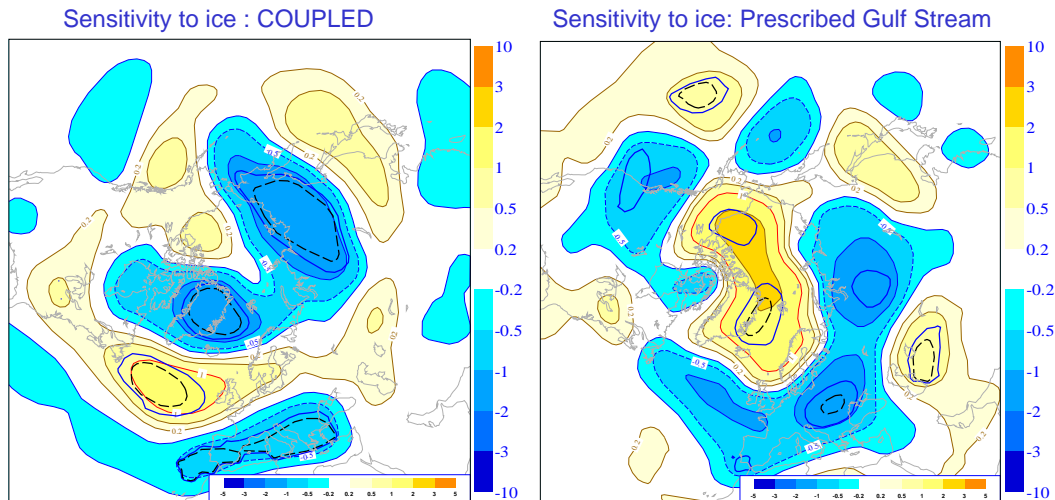


Figure 17: Atmospheric response to the 2008 ice anomaly in the coupled model (left) and in the experiment where the Gulf stream has been corrected (right). Units are dam . The 90% and 95% significance level are shown by the thick blue and dashed-black contours.

model to maintain and propagate the strong MJO (and associated strong westerly wind events) that was present in the initial conditions. Vitart et al. 2006, Woolnough et al. 2006 have shown that a better representation of the mixed layer process is needed improve the predictions of the MJO at the monthly time scales. In particular, they highlight the importance of the diurnal cycle and its rectification in the representation of the intra-seasonal variability. In the case study discussed in the paper, the MJO is in the initial conditions, and therefore, an improved MJO prediction at monthly time scales would result in better seasonal forecast.

The existence of warm bias in the Eastern Pacific seems to be a limiting factor for the amplitude of the SST interannual variability, since the atmospheric response to a heat source is quite non linear. Experiments with different initialization strategies but the same coupled model indicate the initialization influences the bias and in the amplitude of the interannual variability. In S3, the warm bias in the Eastern Pacific and low amplitude of ENSO is a consequence of the initialization. The current initialization strategy, although in general improves the forecast skill by making use of all available information, induces undesirable adjustment processes that can lead to bias and, at times, degradation of the forecast. More balanced initialization strategies are required.

The skill of the seasonal forecasts for the Northern European summers is particularly poor. Although it could well be that the intrinsic predictability of the European summers at seasonal time scales is low, the negative values of the anomaly correlation skill score indicate that model errors are responsible for the negative skill. In this context, some known deficiencies of the coupled model that could affect the poor forecast skill are discussed: ocean mixed layer, Arctic ice forcing and western boundary currents.

The coupled model can not reproduce the rapid shallowing of the ocean mixed layer from spring into summer, which increased the thermal inertia of the ocean, and persists the spring anomalies for too long. This is another example of non linearity in the model, where the a-posteriori correction of the bias is not adequate to remove a systematic error that manifests on a delay response, affecting the timing of the events. The excessive persistence of spring temperature anomalies in the North Atlantic is a common error in the DEMETER seasonal integrations (van Oldenborgh 2007) and needs to be investigated further.

The lack of prognostic sea-ice in the current seasonal forecasting system can also be damaging for the forecasts over Europe, particularly in the last two summers. Atmosphere-only sensitivity experiments indicate that the

ice anomalies in 2007 and 2008 had a significant impact on the atmospheric circulation over the Euro-Atlantic sector, characterized by a high over the Arctic and low centers over Western Europe and North-West America. In the coupled model, however, the sensitivity to the ice anomaly is quite different. Further experimentation indicates that the response of the atmosphere to a given ice anomaly is flow-dependent, being largely conditioned by the background atmospheric mean state. It has been shown that the misrepresentation of the Gulf Stream in the coupled model influences both the mean atmospheric circulation and its sensitivity to the ice anomalies. Whether the Gulf Stream has an impact on other atmospheric tele-connections (such as those caused by tropical forcing) needs to be investigated further.

Although the experiments shown here indicate that the ice anomalies of the past two summers had a significant impact on the atmosphere, the predictability of sea ice anomalies in coupled models is still poorly understood, and it is likely that accurate initialization of sea-ice properties is needed to predict such anomalies a few months in advance.

Finally, most of examples discussed in this work illustrate the limits of the linear assumption implicit in the a-posteriori bias correction used in the production of the seasonal forecasts. The a-posteriori correction of the bias removes only the error in the mean, and it is not sufficient if the systematic error of the model also manifests in other aspects, such as the amplitude of the interannual variability (underestimation of the ENSO amplitude), timing of the events (too much persistence of the spring conditions), or spatial distribution of the anomalies (wrong sensitivity to the ice anomaly).

Acknowledgements

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