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# The relative influence of soil moisture and SST in climate predictability explored within ensembles of AMIP type experiments

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**Abstract** Three ensembles of AMIP-type simulations using the Arpege-climat coupled land–atmosphere model have been designed to assess the relative influence of SST and soil moisture (SM) on climate variability and predictability. The study takes advantage of the GSWP2 land surface reanalysis covering the 1986–1995 period. The GSWP2 forcings have been used to derive a global SM climatology that is fully consistent with the model used in this study. One ensemble of ten simulations has been forced by climatological SST and the simulated SM is relaxed toward the GSWP2 reanalysis. Another ensemble has been forced by observed SST and SM is evolving freely. The last ensemble combines the observed SST forcing and the relaxation toward GSWP2. Two complementary aspects of the predictability have been explored, the potential predictability (analysis of variance) and the effective predictability (skill score). An analysis of variance has revealed the effects of the SST and SM boundary forcings on the variability and potential predictability of near-surface temperature, precipitation and surface evaporation. While in the tropics SST anomalies clearly maintain a potentially predictable variability throughout the annual cycle, in the mid-latitudes the SST forced variability is only dominant in winter and SM plays a leading role in summer. In a similar fashion, the annual cycle of the hindcast skill (evaluated as the anomalous correlation coefficient of the three ensemble means with respect to the ''observations'') indicates that the SST forcing is the dominant contributor over the tropical continents and in the winter mid-latitudes but that SM is supporting a significant part of the skill in the summer mid-latitudes. Focusing on boreal summer, we have then investigated different aspects of the SM and SST contribution to climate variations in terms of spatial distribution and time-evolution. Our experiments suggest that SM is potentially an additional source of climate predictability. A realistic initialization of SM and a proper representation of the land–atmosphere feedbacks seem necessary to improve state-of-the-art dynamical seasonal predictions, but will be actually efficient only in the areas where SM anomalies are themselves predictable at the monthly to seasonal timescale (since remote effects of SM are probably much more limited than SST teleconnections).

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## 1 Introduction

Intraseasonal and interannual atmospheric variability is made of variability internal to the atmosphere, arising from intrinsic chaos and the atmospheric state dependence on its initial conditions, and variability forced by changes at the boundaries, mainly arising from the interactions with the other major climate system components such as the ocean or the land surface. Climate or atmospheric prediction can thus be theoretically decomposed in two distinct problems, an initial value problem and a boundary value problem (Collins and Allen 2002).

Seasonal climate forecasting is physically based on the atmospheric sensitivity to its slowly evolving boundaries. A skillful atmospheric forecast is possible if and only if we are able to understand the atmospheric response to the anomalous boundary forcings and if we are able to anticipate these future boundary forcings. Potential predictability, which can be understood as an upper bound of the real skill, can be estimated by evaluating the atmospheric sensitivity to prescribed boundary forcings (Zwiers 1996; Rowell 1998; Conil and Li 2003). Most of the attention has been given to the SST forcing because of the atmospheric sensitivity to tropical SST anomalies, the so-called teleconnections (Trenberth et al. 1998) providing the basis of the early seasonal prediction (Goddard et al. 2001). Soil moisture (SM) may increase the forecast skill and thus be relevant for seasonal prediction particularly in regions and seasons where and when the atmospheric response to SST forcing is weak (Koster et al. 2000; Douville and Chauvin 2000). The evaluation of the SM based predictability has been set back partly because of the lack of reliable global reanalysis or observations and partly because of the implementation of the land surface models. The SM contribution to the predictability is conditioned by two questions: How predictable are the SM anomalies themselves? How strong is the atmospheric response to SM forcing?

Soil moisture memory is controlled by the seasonality of the atmospheric state, the dependence of evaporation on SM, the variation of runoff with SM and by the coupling between SM and the atmosphere (Koster and Suarez 2001). Wu and Dickinson (2004) have quantified the timescales over which these factors act. SM memory is short within the Tropics and increases with latitude (Manabe and Delworth 1990). It increases with depth and varies with season but it is relatively longer in arid regions. Furthermore SM memory can be much longer in dry conditions than in wet cases in warm climates (Wu and Dickinson 2004).

Previous studies have shown that SM persistence could be translated to atmospheric persistence in a small number of regions (Schlosser and Milly 2002). Atmospheric variability in transition zones between dry and humid areas is particularly affected by land surface forcings (Koster et al. 2000 and 2002). On the other hand, SM predictability was found to be high in the arid/semiarid regions and smaller over rainy temperate zones and tropical monsoon regions in the Kanamitsu et al. (2003) experiments.

Two types of numerical experiments have been conducted to assess the SM influence on climate variability focusing either on the initial or boundary conditions problem. In the context of dynamical seasonal prediction, many studies have been devoted to the impacts of SM initialization on forecast skill. Koster et al. (2004a) have shown a small increase of the skill in the Great Plains in summer. Fennessy and Shukla (1999) concluded that the influence of realistic initial SM was mainly local and dependent on several factors including the area extent and magnitude of the initial SM anomaly, its persistence, and also the mean climate of the region. While they demonstrated the incompatibility between observed and modeled SM, Zhang and Frederiksen (2003) have indicated that realistic relative SM anomalies may contain useful informations for improving the simulated climate mostly locally.

Douville (2004) has suggested that the atmospheric variability is sensitive to the SM feedbacks while atmospheric potential predictability is mostly sensitive to the initial SM. In most studies, realistic initialization of SM has led to mixed results concerning the skill of precipitation or temperature forecast as in Douville and Chauvin (2000), Koster and Suarez (2003b) or Dirmeyer (2003). The design of the experiment itself can affect the atmospheric response to land surface forcing (Douville 2003, 2004).

A small number of potential predictability studies based on different experimental design and SM data have led to contrasting results. Yang et al. (2004) designed an ensemble of AGCM forced simulations using observed SST and ‘‘reanalyzed’’ SM over the period 1979–2000. Focusing their analysis on the US region and the summer period, they found small (negligible) near surface temperature (precipitation) predictability. Using high-resolution SM climatologies for the contrasting years 1987 and 1988 derived from the first global soil wetness project (GSWP), Dirmeyer (2000) and Douville and Chauvin (2000) have shown the relevance of realistic SM for simulating seasonal precipitation.

The use of ‘‘realistic’’ SM also enhanced hindcast scores and potential predictability over semi-arid regions in the GCM experiments designed by Kanae et al. (2006). In a perfect model framework (GLACE project), Koster and the GLACE team (2004b) pointed out such semi-arid regions as climate ‘‘hot spots’’ where the land–atmosphere coupling is particularly strong. Dirmeyer (2005) assessed the predictability gained by different combinations of flux replacements and initial conditions. While land initial conditions helped to create realistic climate anomalies in some regions, the flux replacements were necessary to maintain the simulated SM in realistic regimes.

This work's objectives are to quantitatively assess the relative contributions of SM forcings to climate predictability. SM forcings are derived from the updated ‘‘realistic’’ GSWP2 SM archive for the period

1986–1995 and original numerical experiments have been designed to combine SST and SM influences. Three ensembles of GCM simulations are used to evaluate the contribution of SST and SM to the potential predictability and to the hindcast skill. Section 2 is dedicated to the presentation of the experiments and to a brief evaluation of the model used. The annual cycles of the potential predictability and hindcast skill are discussed in Sect. 3. Section 4 focuses on the summer mid-latitudes. In this section we particularly investigate the relative influence of SM and SST on two case studies over the North American and European regions. A summary and discussion of the results is given in Sect. 5.

## 2 Ensemble simulations

### 2.1 Design of the experiments

The present study is devoted to the analysis of the respective role of SST and SM in maintaining climate variability and predictability, with an emphasis on the role of realistic low frequency SM forcing. GSWP2 has offered the opportunity to produce a land surface reanalysis (state-of-the-art global data sets of land surface fluxes, state variables, and related hydrologic quantities, including soil moisture) over the 10-year period from 1986 to 1995, fully consistent with ISBA, the land surface model (LSM) currently used in the Arpege-Climat model (Déqué et al. 1994; Douville 1998). Note however, that the SM data (referenced as the ISBA GSWP2 analysis in the following) used in this study is not exactly the ISBA climatology available from the GSWP2 database, which was obtained with a 3-layer rather than a 2-layer soil hydrology as is currently coupled to the Arpege GCM, and with the International Satellite Land-Surface Climatology Project (ISLSCP) Initiative II (Hall et al. 2003) rather than ECOCLIMAP (Masson et al. 2003) soil and vegetation parameters.

Though certainly model dependent, the ISBA GSWP2 SM climatology used in the present study is consistent with other related or independent datasets. Figure 1 illustrates the relative agreement between the SM and latent heat flux (LHF) derived from the 2-layers ISBA LSM, the GSWP2 multi model ensemble mean (Dirmeyer et al. 2005) and the ERA40 reanalysis (Uppala et al. 2005). The upper panels show the grid point anomaly correlation coefficient (ACC) between the ISBA SM and LHF, and the multi-model GSWP2 reanalysis. The middle (lower) panels compare the time evolution of the Great Plains (NE Europe) regional average SM and LHF in the three reanalyses. The global average grid point correlation between the monthly ISBA SM (LHF) and the multi-model GSWP2 analysis over the 10 years is 0.71 (0.69), but most of the grid points show much higher correlations. A relatively small number of grid points, mainly located in very dry areas where anomalies are almost insignificant or in permafrost regions where the disagreements between the LSMs are the largest, show very low correlations. The Great Plains area has experienced large variability during the 1986–1995 period. The multi model and the ISBA GSWP2 analysis SM are in a good agreement with the ERA40 even if some disagreements are found during the 10-year period. As also evidenced in the ACC map, the latent heat flux anomalies in the three reanalyses have weaker correspondences over the summer mid-latitudes.

We used the ISBA GSWP2 land surface analysis to explore the respective role of SST and SM in sustaining climate variability and predictability. The SST are used as raw boundary conditions for the atmospheric GCM as commonly used for evaluating the upper limit of seasonal predictability (Gueremy et al. 2005). Realistic SM were imposed by a relaxation of the total SM toward the ISBA GSWP2 monthly reanalysis allowing control of the low frequency variations of total SM, the potential major contributor to climate predictability due to its longer persistence, while leaving the atmosphere and the land surface free to interact at high frequency. Recall that the original ISBA scheme is based on a two-layer hydrology with the surface reservoir included in the total soil moisture reservoir. Further technical details on the nudging can be found in Douville (2003).

The Arpege AGCM is used to perform global simulations in which the SM is either fully interactive or relaxed toward the GSWP2 analysis. Three AMIP type-experiments composed of ten members have been performed over the 10-year period 1986–1995. In the first ensemble EE, the SM is relaxed toward the GSWP2 analysis and the SST is kept climatological.

In the second ensemble FF, the observed SST is used to force the AGCM and the SM is fully interactive. In the last ensemble GG, the SM is relaxed toward GSWP2 as in EE and the SST are observed as in FF.

| Experiment's name | SST | SM |
|---|---|---|
| EE | Climatological | Relaxed toward GSWP2 |
| FF | Observed (AMIP) | Fully interactive |
| GG | Observed (AMIP) | Relaxed toward GSWP2 |

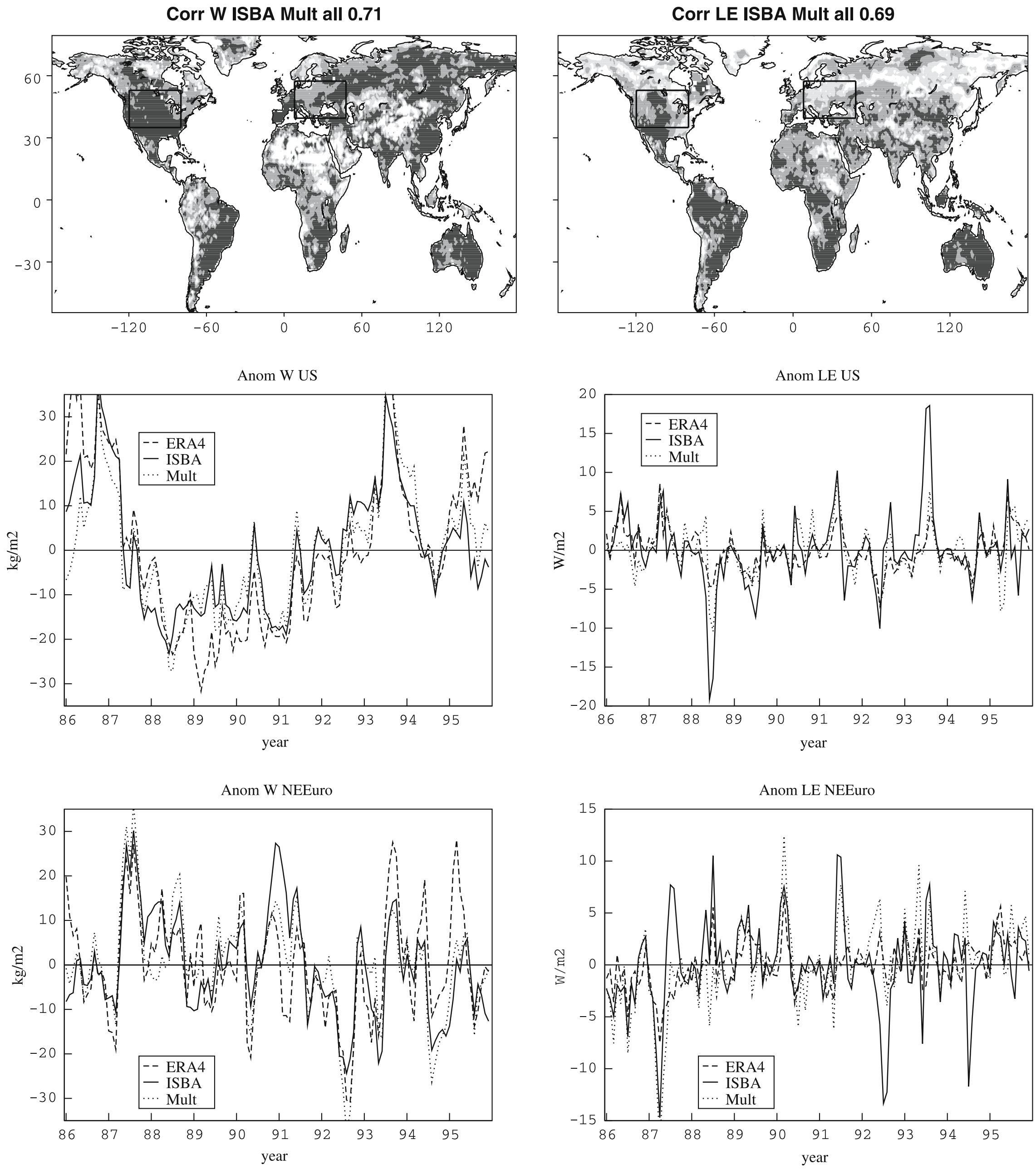


**Fig. 1** Relative agreement between ISBA and the Multi Model GSWP2 and ERA40 reanalysis. Soil moisture (latent heat flux) is considered on the *left* (*right*) panels. Grid point temporal anomaly correlation coefficient between ISBA and multi model GSWP2 analysis are shown on the *upper panels* over the entire 1986–1995 period. The 0.4, 0.6 and 0.8 levels are shaded in *light* to *dark greys*. The two black boxes are defining the US and NE Europe regions used later in the analysis. Time series of the soil moisture and latent heat flux anomalies averaged over the US (NE Europe) region are presented in the *middle* (*lower*) *panels*.Correlation of ISBA and multi model GSWP2 (ERA40) soil moisture anomalies reaches 0.86 (0.64) over the NE Europe and 0.91 (0.85) over the US. Correlation of ISBA and multi model GSWP2 (ERA40) latent heat flux anomalies reaches 0.59 (0.42) over the NE Europe and 0.66 (0.60) over the US. Units: kg $m^{-2}$ and W $m^{-2}$

The two perturbed experiments are not truly symmetrical, but SST and SM are not symmetrical variables. First of all, we have assumed that the SST anomalies might exert a large influence on the SM anomalies through the associated large-scale teleconnections while the influence of SM on the SST anomalies is supposed to be much weaker. Furthermore, in most of the AGCMs, SSTs are imposed as boundary conditions but SM is fully prognostic and takes part in the complex land–atmosphere interactions. The control experiment, nicknamed FF, combines realistic ‘‘observed ‘‘ SST and fully interactive SM. The two perturbed experiments are designed to highlight the influence of realistic SM, in two different contexts, one with realistic ‘‘observed’’ SST, GG, and one with climatological SST, EE.

### 2.2 Validation

As already shown by Douville and Chauvin (2000) and Douville (2003) the relaxation of SM toward the GSWP analysis can lead to some improvements in the model climatology. The climatology of the two relaxed SM experiments EE and GG are almost identical, their differences being mostly indiscernible. The interannual variations of SST do not perturb the mean state but modifiy its variability. To assess the influence of realistic SM on the simulated climate we have thus compared the experiment FF (our control experiment) and the experiment GG, which is relaxed toward realistic SM. The relaxation of the SM toward the GSWP2 SM analysis has a mostly positive impact on the model climatology particularly in the summer hemisphere.

The differences between the simulated evaporation, precipitation and 2 m temperature climatology (ensemble mean from the FF experiments) and the reanalysis (GSWP2 for the evaporation and ERA40 for the 2 m temperature) or a rain gauge analysis, GPCC, for the precipitation, are shown for the summer season (JJAS) on the left panels of Fig. 2. Hagemann et al. (2005) have assessed the realism of the hydrological cycle of ERA40 and Zhao and Dirmeyer (2003) have evaluated the GSWP2 data. On the right panels the differences between the relaxed SM experiment (GG) and the control experiment FF during summer are displayed. Most of the large-scale features of the land surface characteristics (not shown) are fairly well captured by the Arpege model including the latitudinal evolution of SM. Nevertheless several systematic errors of the SM regional distributions are noticeable. The mid to high summer (northern) latitudes are too wet while the tropics and subtropics seem mostly too dry. Strong biases are also found in the Indian monsoon region. While such biases are quite common in land–atmosphere modeling, they are perturbing the variability of the land–atmosphere system because of the intrinsic relations between mean state and variability. These perturbations are problematic since they obscure the predictability of the climate, particularly in summer when the oceanic influence is weaker.

The systematic errors in the simulated SM climatology are entirely corrected by the relaxation (not shown). These SM biases are intimately related to biases in the surface water budget and particularly to precipitation and evaporation and thus systematic errors of precipitation and evaporation are strongly collocated with the SM ones. As also shown by the COLA model (Dirmeyer and Zhao 2004) and the NASA model (Koster et al. 2000), the Arpege model overestimates the precipitation in the northern mid to high latitudes. Rainfall in the monsoon regions is also not well simulated. While SM is clearly highly sensitive to mean precipitation, it should be kept in mind that mean precipitation also depends on SM. Indeed, the relaxation of SM toward a realistic analysis reduces (but does not suppress) the model systematic errors of precipitation and evaporation (see right panels of Fig. 2). On the other hand, while the biases of atmospheric temperature and pressure (not shown) are affected by the GSWP2 SM forcing, they are not however, systematically diminished. This suggests that, while the SM biases have contributed to these near surface temperature biases, the complex large-scale atmospheric thermo-dynamical equilibrium is responsible for much of these systematic errors. In fact the simulated lower troposphere temperatures are too warm over much of the northern continents and the meridional pressure gradient is weaker than in the reanalysis.

The use of realistic SM conditions may thus improve the climatology of the simulated climate when the systematic errors are related to the land–atmosphere coupling. Conversely when the biases are more likely controlled by the vertical/latitudinal structure of the simulated atmosphere, e.g., the large-scale thermo-dynamical equilibrium, the SM relaxation, which still has noticeable impacts on the climatology, might not lead to an overall and systematic improvements. Some improvement of the model climatology may thus be achieved through the use of SM relaxation, mainly the characteristics highly sensitive to the land–atmosphere interactions.

## 3 Annual cycle of potential and effective predictability

### 3.1 ANOVA

In order to quantify the effects of SM and SST on climate variability we make use of a one-way analysis of variance (ANOVA) for which the previously described simulations were designed. The ANOVA is a powerful tool allowing the decomposition of the variance and thus the estimation of the variability forced by SM and SST and its ratio over the total variability, often referred to as potential predictability (PP), in the perfect model approach. Details on the methodology

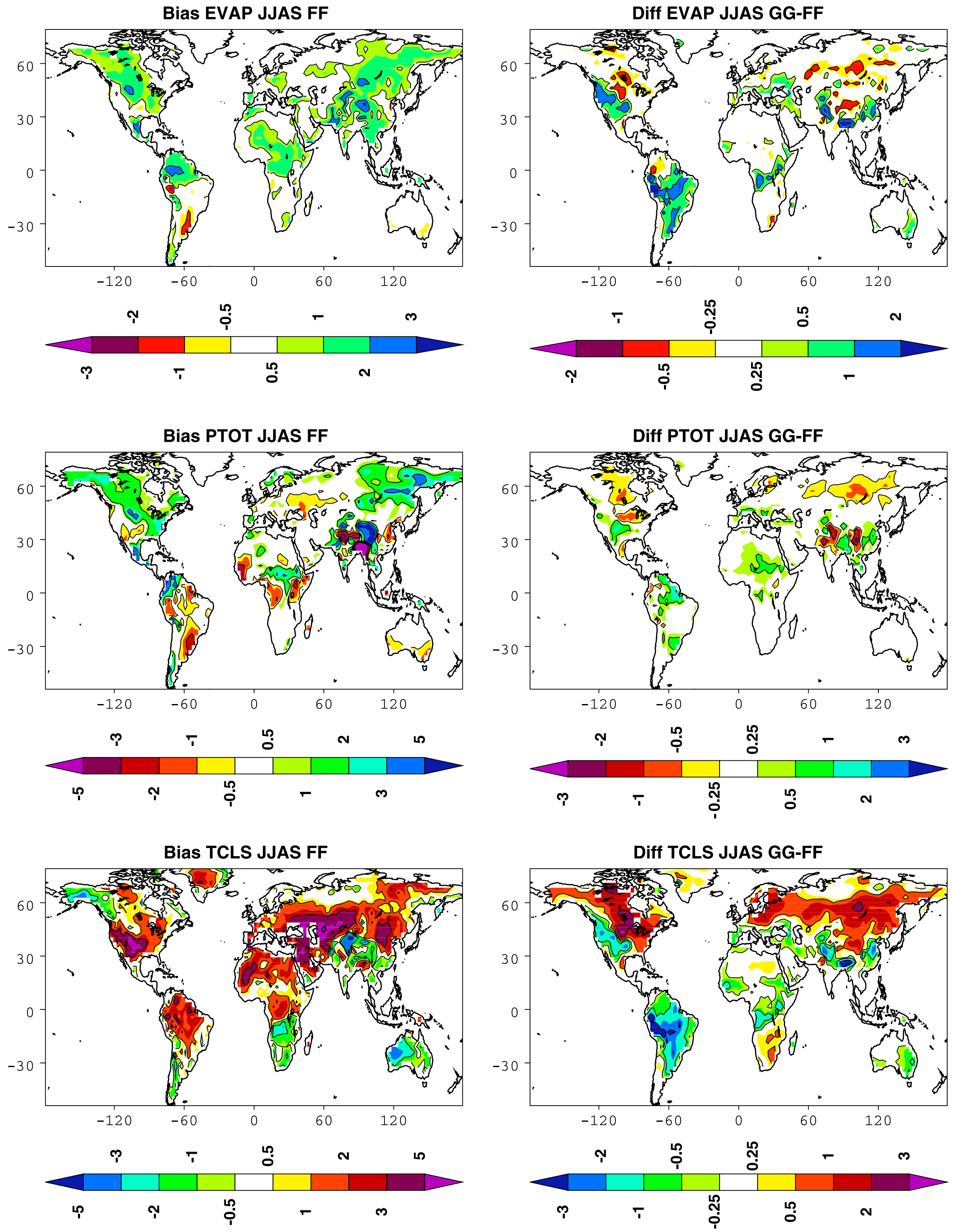


**Fig. 2** Bias of the standard Arpege climatology (*left panels*) and impacts of the SM nudg- ing on the Arpege climatology (*right panels*) during summer. Evaporation, precipitation and 2 m temperature are shown respectively on the *upper, middle* and *lower panels*. Evaporation bias is the difference between the mean summer evaporation in the FF experiment and the GSWP2 reanalysis. Precipitation and 2 m temperature biases are evaluated aginst the GPCC data and the ERA40 reanalysis. The effects of realistic SM on the Arpege climatology is shown as the difference between the summer (JJAS) climatology estimated in the GG and FF ensembles (see text for details). Units: mm day$^{-1}$, mm day$^{-1}$ and K

and its underlying hypotheses can be found in Von Storch and Zwiers (1999).

In the experiments EE and GG, while the fast land–atmosphere interactions are free, the low frequency variations of SM are imposed by a nudging with a one-day timescale.

In these two experiments, the interannual variations of monthly mean SM are thus almost entirely constrained, the forced variability (that can be attributed to the boundary conditions) almost equals the total variability and thus the potential predictability of SM is almost 100%. The SM variability forced by the SST and its potential predictability are higher in the equatorial region and then decrease toward the poles. Minimum predictability (less than 5%) is found in the mid-latitudes during summer while the equatorial band maximum predictability reaches 50% during summer.

The annual cycles of the zonal mean forced variability and PP of evaporation (precipitation) are shown in Fig. 3 (Fig. 4). Precipitation and evaporation over land are partly constrained by the boundary forcings. While the SM forcing exerts a greater control over evaporation (and precipitation) in the northern mid-latitudes during summer, the SST forcing is stronger in

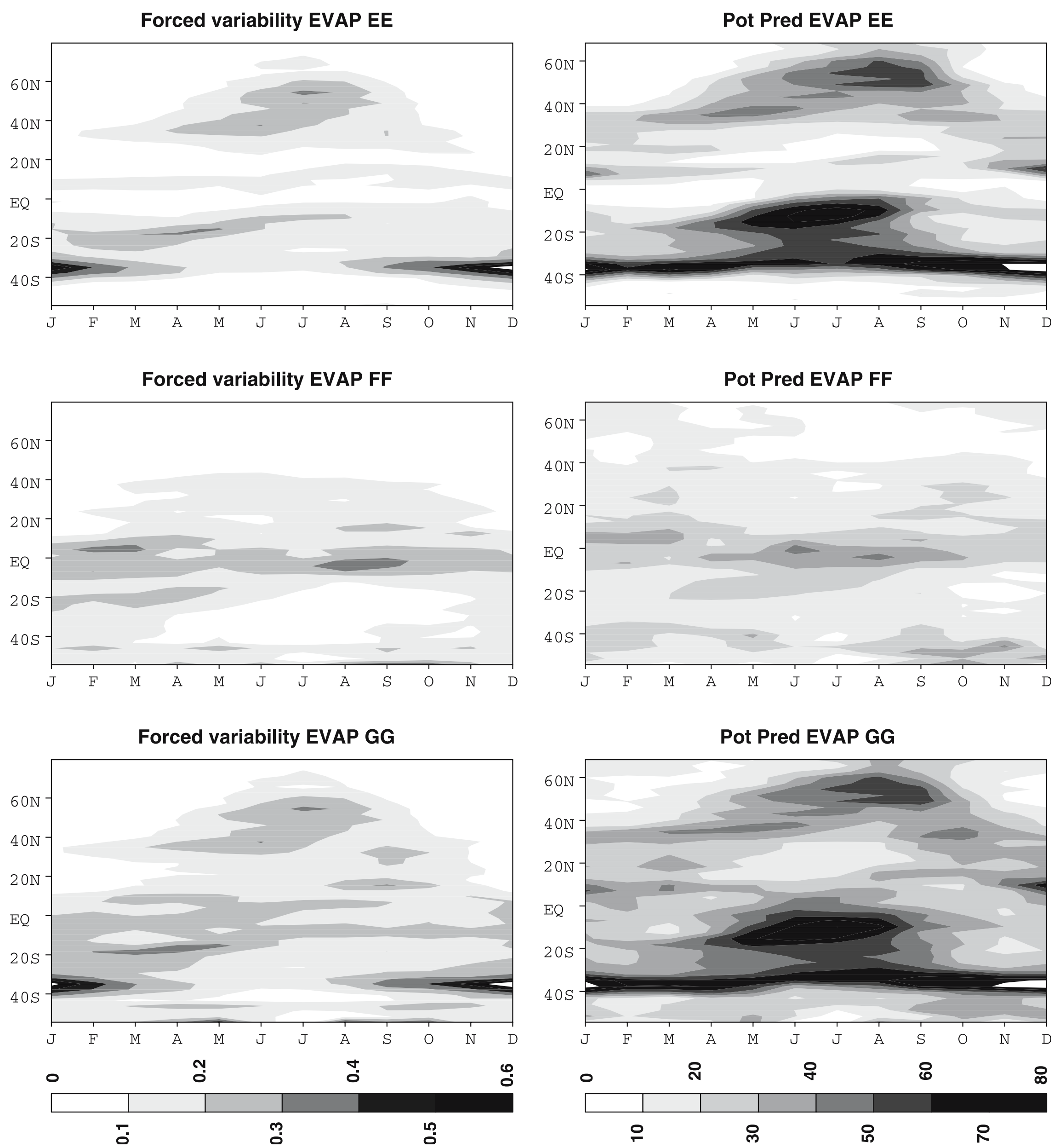


**Fig. 3** Annual cycle of the zonal mean forced variability (*left panels*, unit: mm day$^{-1}$) and potential predictability (*right panels*, unit: %) of evaporation in the three ensembles EE, FF and GG, evaluated over continents

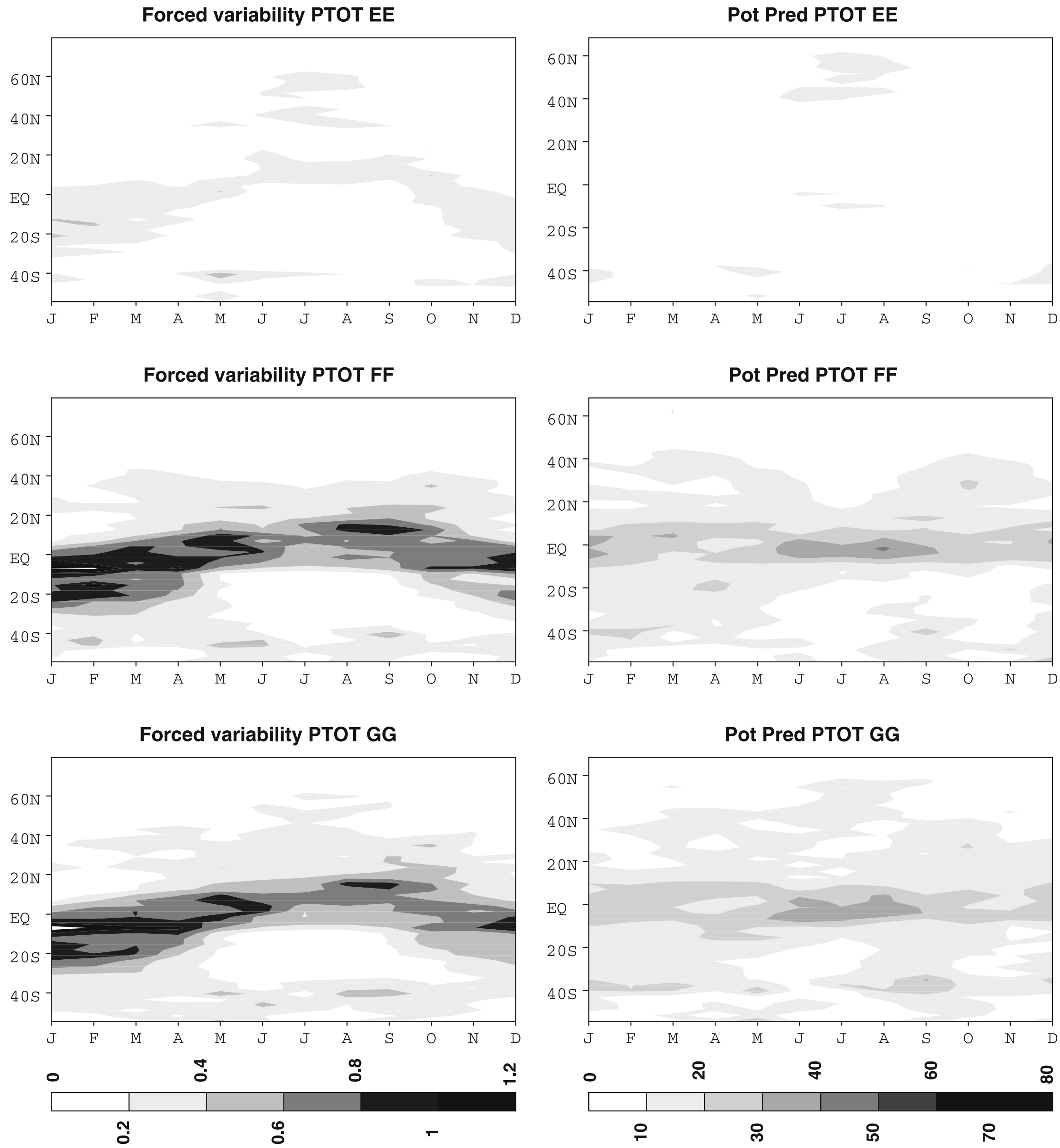


**Fig. 4** Annual cycle of the zonal mean forced variability (*left panels*, unit: mm day$^{-1}$) and potential predictability (*right panels*, unit: %) of precipitation in the three ensembles EE, FF and GG, evaluated over continents

the tropical region (particularly for the precipitation). The SST (SM) conditions contribute to the greatest part of the precipitation (evaporation) forced variability in the tropical (mid-latitudes) regions.

As seen in the GG experiment, SST and SM boundary conditions are playing two complementary roles in the maintenance of the PP of evaporation and precipitation. While precipitation variability forced by SM is weaker, it still contributes to the potential predictability in the northern mid-latitudes during summer. On the other hand, the SM boundary conditions are the major contributor to the evaporation potential predictability.

In the extratropics during summer, while some 50–60% of the evaporation variations are potentially predictable, most of the precipitation variance remains of chaotic nature.

SM, evaporation and precipitation have their maximum PP collocated with their maximum forced variability. But air temperature and pressure have their maximum potential predictability in the tropical regions (where variability is small) while they have their maximum forced variability in the mid and high latitudes (where potential predictability is rather small). Forced variability and PP of 2 m temperature estimated in the three experiments are shown in Fig. 5. Similar patterns of seasonal evolution of lower/mid troposphere temperature and sea level pressure (SLP) are obtained but with weaker amplitude (potential predictability of SLP and 850 hPa temperature reach 20–30% and 30–40% in the mid-latitudes during summer in the experiments EE and GG).

In the tropics, the PP of atmospheric temperature and pressure is almost entirely explained by the SST boundary conditions, all year round. In the extratropics, the SST boundary conditions are the main contributor to the temperature and pressure forced variability during winter while the SM boundary conditions are taking the leading role in summer. The internal chaotic variability of temperature and SLP is strong in winter, explaining a low potential predictability during this season. But during summer, the anticipated SM boundary conditions are able to force most of the variability and thus to maintain a significant potential predictability. Figure 6 shows the summer (JJAS) potential predictability of evaporation,

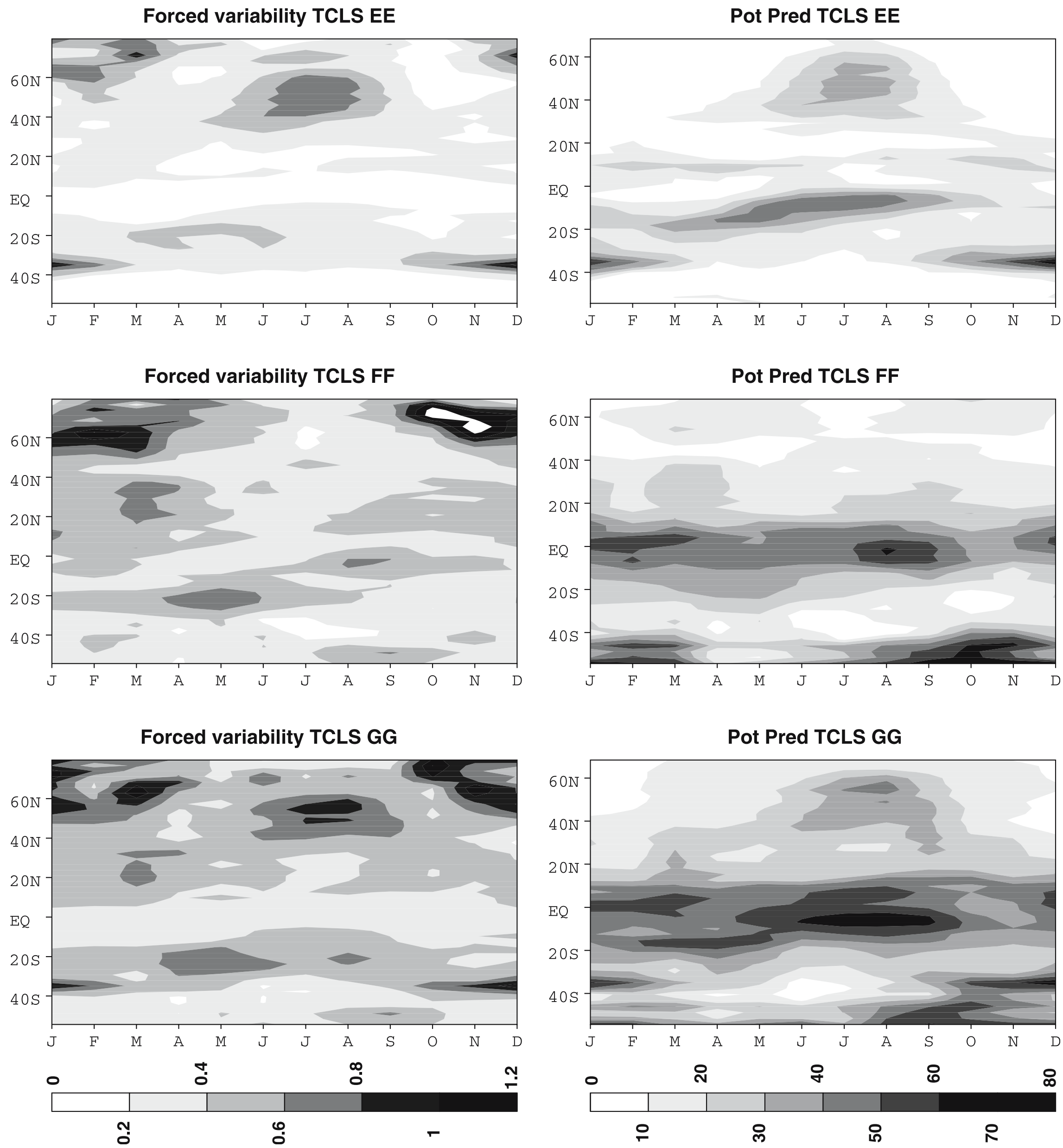


**Fig. 5** Annual cycle of the zonal mean forced variability (*left panels*, unit: K) and potential predictability (*right panels*, unit: %) of 2 m temperature in the three ensembles EE, FF and GG, evaluated over continents

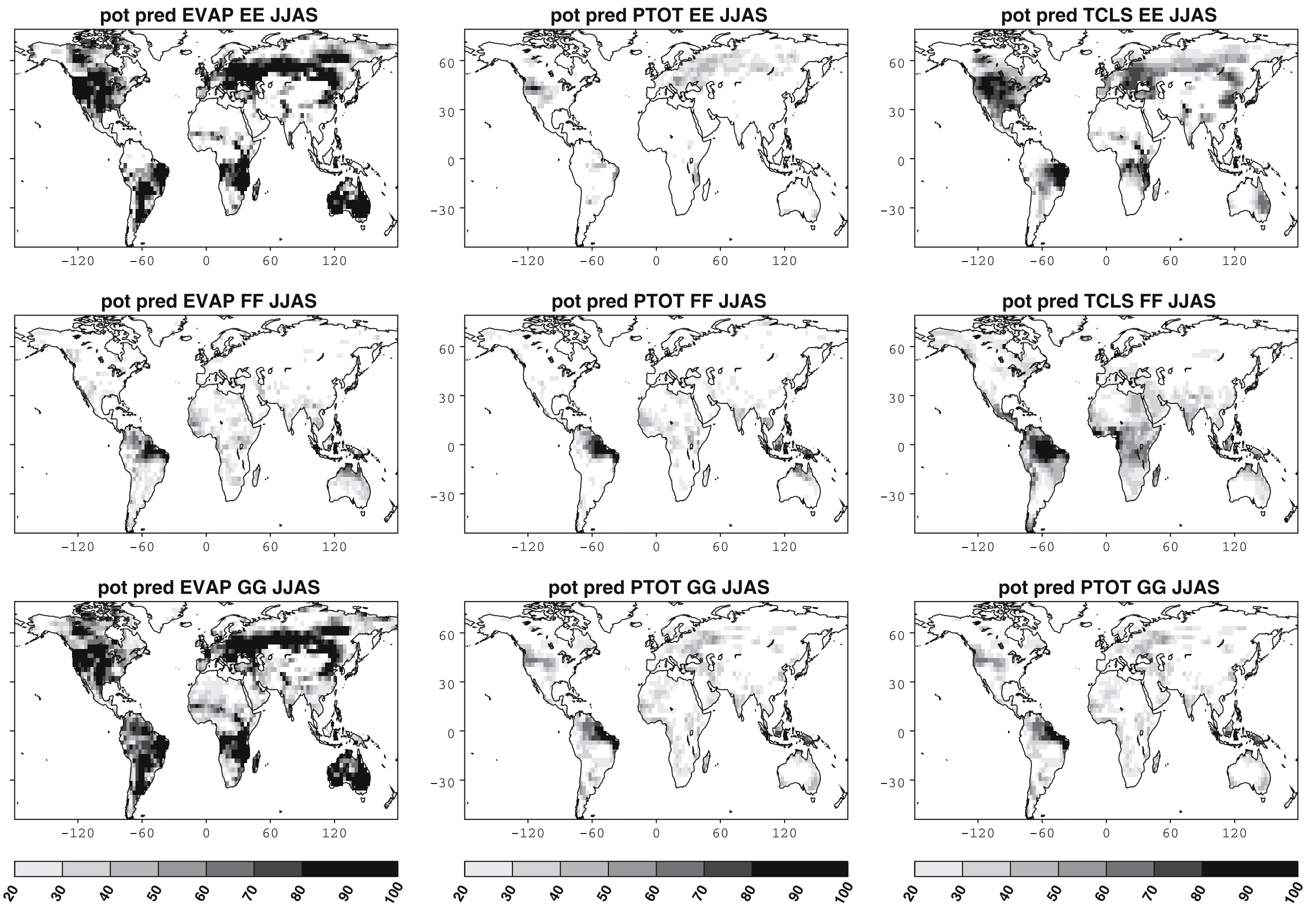


**Fig. 6** Summer (JJAS) potential predictability of evaporation (*left panels*), precipitation (*center panel*) and 2 m temperature (*right panels*) estimated in the three ensembles EE (*upper panels*), FF (*middle panels*) and GG (*lower panels*)

precipitation and near-surface temperature estimated in the three ensembles. In the summer season, evaporation and temperature are more likely controlled by land (and secondly ocean) boundary conditions, precipitation being of a more general chaotic nature. The SST influences predominantly the climate variations in the tropical regions while the SM controls most of the near surface climate changes in the mid-latitudes. Koster et al. (2000) using similar sets of experiments also concluded that the relative contributions of SST and SM forcings to summer precipitation PP are clearly separated, the land–atmosphere feedback being the most important outside of the tropics which are mainly influenced by SST. Kanae et al. (2006) also revealed that summer PP of precipitation and evaporation were improved by the use of realistic SM forcing over some mid-latitudes continental areas. In their ensembles of simulation, the influence of SM on summer precipitation and evaporation PP have a much smaller spatial extension in the mid latitudes, particularly over the Eurasian continent. But in the idealized experiments designed by Koster et al. (2000), the SM forcing exerts a significant influence on summer precipitation PP over large parts of the Eurasian continent.

## 3.2 Effective predictability (hindcast skill scores)

While the previous section was dedicated to the estimation of the part of the climate variations that can be attributed either to SM or SST boundary conditions, here we will quantify the ability of the model to reproduce the observed past climate variations, e.g., the skill of the model, by calculating the grid point temporal anomaly correlation coefficient (TACC) between observed and simulated anomalies of evaporation, precipitation and near surface temperature in the three experiments EE, FF and GG. The design of our experiments, mainly the use of realistic SM and SST conditions, allow us to verify our simulations against the observations and thus to estimate if the simulated forced variability is realistic. Even if the model were perfect, precipitation would have a weak ACC since it is more of a chaotic nature while evaporation and temperature seem more predictable and thus should reach higher scores. The ‘‘observations’’ used for the skill estimation are derived from the ERA40 reanalysis (2 m temperature), from the GSWP2 reanalysis (evaporation) and from the GPCC rain gauge analysis (precipitation).

The seasonal evolutions of the zonal average of precipitation and 2 m temperature TACCs in the three experiments EE, FF and GG are displayed in Fig. 7. As expected, globally, temperature TACCs are higher than precipitation TACCs, temperature being easier to forecast than precipitation, particularly with prescribed SST. The SM forcing helps to maintain good temperature and precipitation skills during summer in the northern mid-latitudes while the SST conditions mainly contribute to the skill in the tropical regions and in the northern mid-latitudes during winter. The combination of the two forcings revealed in the GG experiment is mostly linear, the temperature and precipitation skills being almost a linear addition of the contribution from SM (estimated from EE) and the contribution from SST (estimated from FF).

Kanae et al. (2006), in similar ensembles of experiments, have found that adequate land surface boundary conditions could enhance the predictability of precipitation on an interannual basis. Dirmeyer and Zhao (2004) and Dirmeyer (2005) examined the role of land surface in climate predictability using a flux replacement method in order to eliminate the climate drift they noticed in their model, but also to obtain realistic land surface forcing. Their summer precipitation skill was low, but showed signs of improvement with flux replacement. Conversely the skill of summer temperature forecasts was fairly good, and improved markedly with flux replacement. Our results and the papers discussed above are suggesting that SM boundary conditions are a major contributor to the

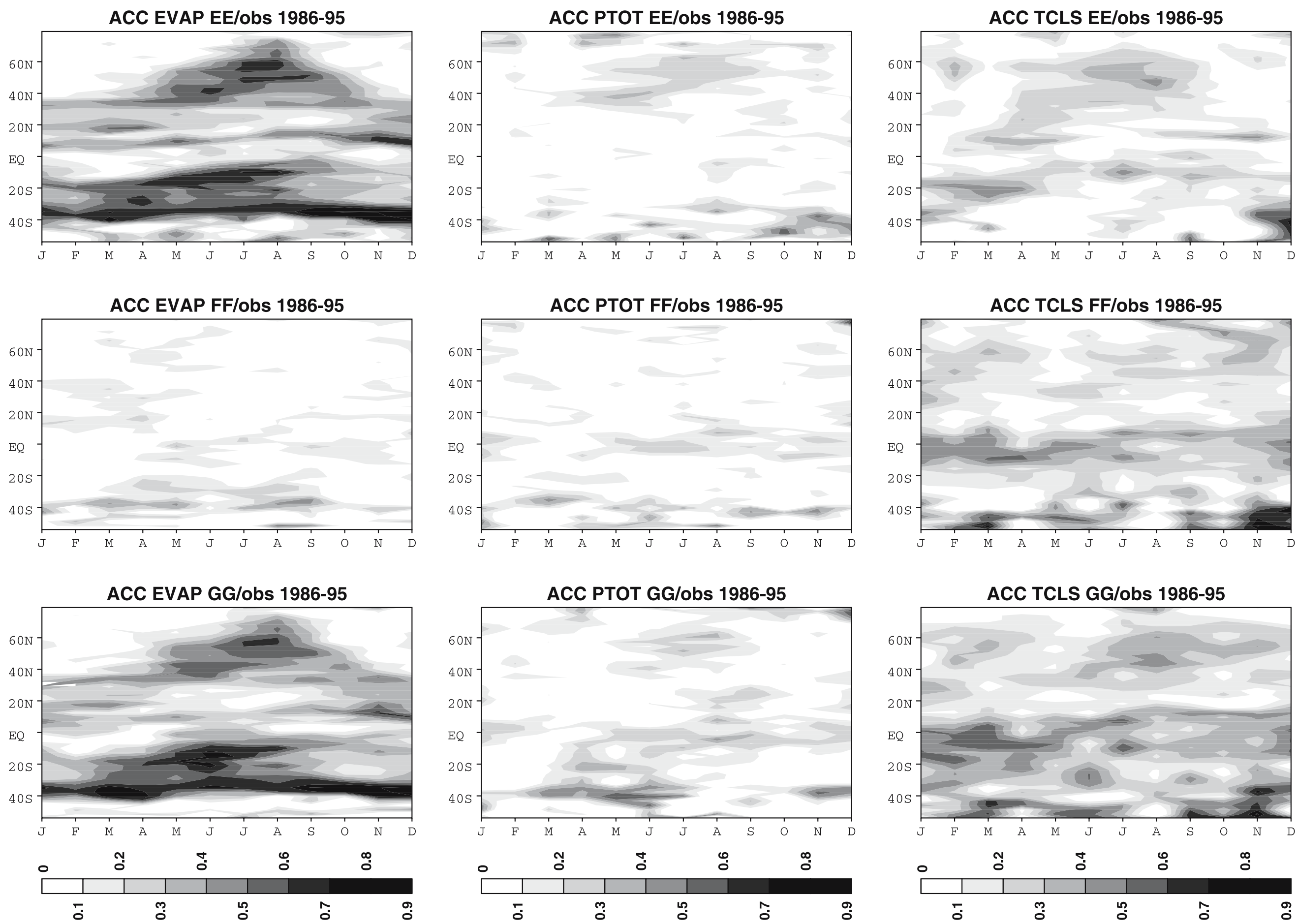


**Fig. 7** Annual cycle of the zonal mean skills (correlation coefficient of the simulated anomalies with the GSWP2 evaporation, the GPCC precipitation and the ERA40 2 m temperature) of the ensemble mean evaporation (*left panels*), precipitation (*center panels*) and 2 m temperature (*right panels*) in the three ensembles EE (*upper panels*), FF (*middle panels*) and GG (*lower panels*). Land points only are considered

climate variations during the summer in the mid-latitudes. They control a large part of the near surface temperature, evaporation and precipitation variations

## 4 Focus on the summer mid-latitudes

### 4.1 Boreal summer skill

Outside from the tropics, where the skill, as well as the variability, has a weak annual cycle, the maximum skill is found in the mid-latitudes during summer. The spatial distributions of the summer evaporation, precipitation and 2 m temperature skills in the three experiments are shown in Fig. 8.

In the EE and GG experiments, high skill scores (> 0.8) are obtained in the northern extratropics for both evaporation and temperature, particularly over the Eurasian continent. Precipitation skills are much lower, but still large parts of the Eurasian and North American continents are above 0.5. Conversely, the skills maintained by the SST forcing only are small except for 2 m temperature in the central US. In the southern winter hemisphere, the SM forcing maintains high skill scores of evaporation, temperature (and precipitation) over tropical South America, Africa and Australia.

The combined effects of SM and SST add almost linearly except at a small number of grid points. While SM almost always adds skills in the northern extratropics, the introduction of the SST forcing can lead to a reduction of the skill, most likely due to a misrepresentation of the atmospheric teleconnections. Nevertheless most of the skill in summer is limited to the lower troposphere. The skill of atmospheric temperature decreases when altitude increases (not shown). Pressure and geopotential skills are high in the tropics, but are small over the mid-latitude continents and are maintained almost entirely by the SST forcing except over North America (not shown).

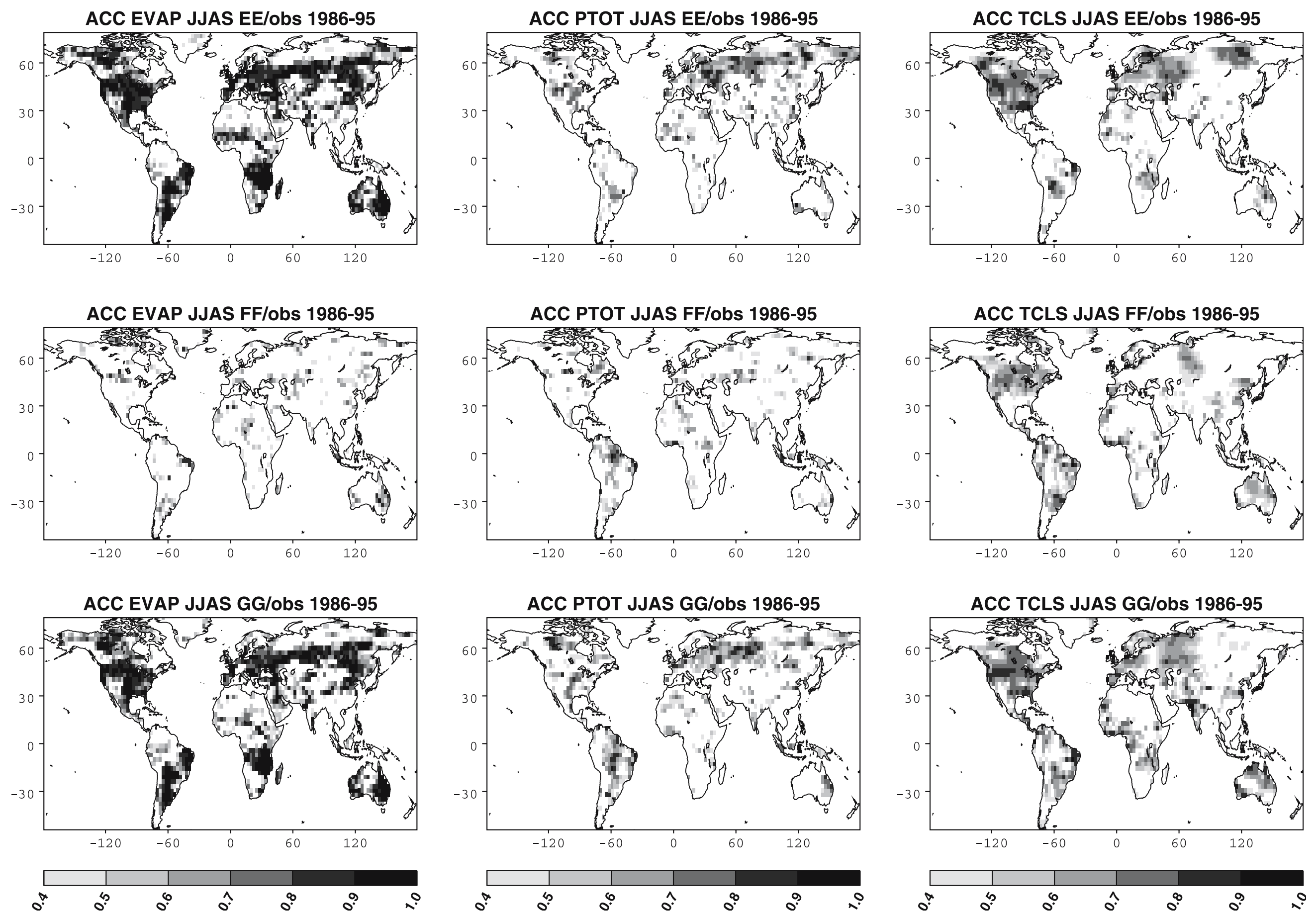


**Fig. 8** Summer (JJAS) skills (correlation coefficient of the ensemble mean anomalies with the ISBW GSWP2 evaporation, the GPCC precipitation and the ERA40 2 m temperature) of the ensemble mean evaporation (*left panels*), precipitation (*center panels*) and 2 m temperature (*right panels*) in the three ensembles EE (*upper panels*), FF (*middle panels*) and GG (*lower panels*)

With the knowledge of the SST and SM states, the atmosphere is able to maintain a potentially predictable signal in precipitation and near surface temperature mainly over the mid-latitude continents during summer (see previous section). The SM forcing leads to a local or regional atmospheric response while the SST forcing is able to influence the atmosphere locally but also to generate remote teleconnections. While local/regional atmospheric responses are generally well represented in the GCM, the large-scale atmospheric teleconnections might be misrepresented leading to a poor hindcast skill but a high potential predictability. In order to be able to achieve skillfull dynamical seasonal forecasts, it is essential to combine reasonable atmospheric sensitivities to its boundary forcings (e.g., a certain amount of potential predictability) and a realistic representation of the atmospheric teleconnections (maintaining good hindcast skills).

Figure 9 shows the 10-year evolution of the spatial ACC (SACC) of temperature and precipitation (simulations against observations) in the northern Hemisphere mid-latitudes (30 N:75 N) over land only. We have computed the SACC of every single simulation and of the ensemble mean. For clarity, we have shown the ensemble mean SACC and a measure of the spread around the mean, the standard deviations of the 10 single SACCs in the three experiments. In the legend, the *m* value is the 10-year average ensemble mean SACCs. The *s* value, describing the mean spread, is the 10-year average of the standard deviations shown on the plots (it is the average uncertainty on the mean SACC values).

The average temperature SACCs are higher than the precipitation SACCs, another illustration of the fact that temperature is easier to forecast than precipitation, on average. The large interannual variations of the SACC suggest that the ability to simulate temperature and precipitation anomaly patterns is not stationary but case dependent. Although most of the skill in reproducing the spatial patterns of the temperature and precipitation anomalies is due to the SM forcing, the improvements resulting from the combination of realistic SST and SM forcings is also case dependent, particularly if you consider precipitation.

For temperature and precipitation, the average spread of the SACC is not modified by the introduction of realistic and variable SM or SST forcings. The noise component of the variability has similar amplitude in the three experiments (as also revealed by the ANOVA analysis, see Sect. 3). The spreads in the temperature SACC are also case dependent. The comparison of 1993 and 1989 shows that while experiencing a similar level of ensemble mean SACC in the three experiments, the summer of 1989 (1993) was poorly (strongly) constrained by the boundary conditions and strongly (weakly) sensitive to the initial conditions. We are now going to illustrate the SM and SST roles in the generation of climate anomalies, focusing on two mid-latitudes regions.

## 4.2 Two cases over North American and European regions

We have selected two mid-latitudes regions of particular interest, North America and Europe, to explore

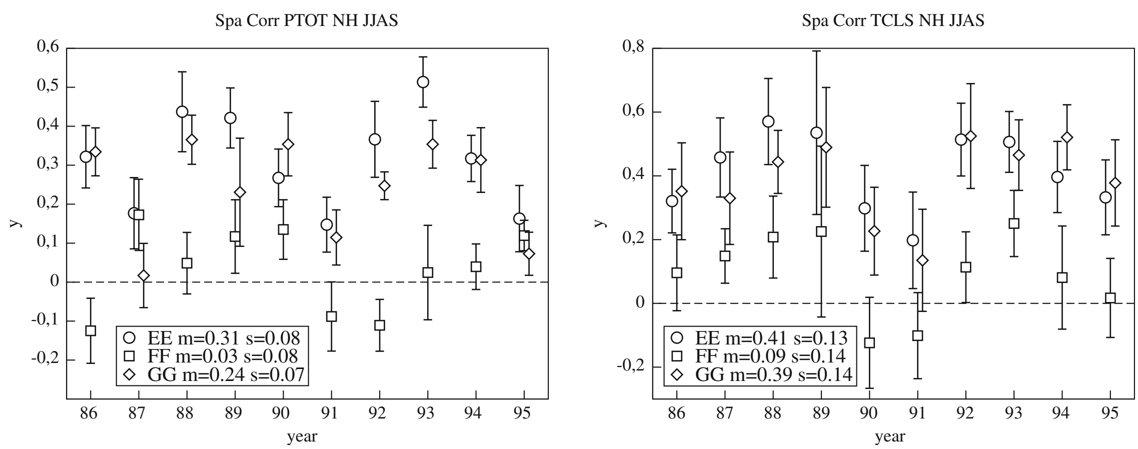


**Fig. 9** Summer (JJAS) Spatial Anomaly Correlation Coefficient (SACC) computed over the continental northern extratropics (30 N/75 N). Precipitation SACC (*left panel*) and 2 m temperature SACC (*right panel*) were estimated from the GPCC rain anomalies and ERA40 2 m temperature anomalies. The SACC of the ensemble mean is marked by a symbol for each summers. The error bars are showing the standard deviation of the SACC in each ensemble. In the legend the average SACC over the ten summers is given by the m value while the *s* value is the average SACC standard deviation

the relative effects of SST and SM boundary conditions on their summer climate. The two regions have experienced strong climate anomalies in the past with dramatic socio-economical consequences. Furthermore, using global correlation maps (not shown), a preliminary analysis of the spatial extension of precipitation and temperature anomalies observed over the period 1960–2000 has revealed that these two regions are characterized by well organized climate and SM anomalies.

The 10-year evolution of precipitation and 2 m temperature anomalies in the three experiments and in the observations are shown in Fig. 10 for the two regions over North America and Europe (the extent of the two regions can be seen on Fig. 1). The ensemble mean is shown by a symbol and the uncertainties by the error bars showing the standard deviations of the anomalies within the ensemble of simulations for a given summer season. The legend contains several statistics useful for the interpretation. The *a* value gives the mean amplitude of the ensemble mean time series, the *r* value gives the correlation coefficient of the ensemble mean and the observed time series, e.g., the skill score, the *s* value is the average uncertainty for the given ensemble, e.g., the average standard deviation within an ensemble. For the temperature evolution as well as for precipitation, in the two regions, the lack of SM forcing significantly reduces the amplitude of the signal (see values of amplitude *a* in the legend of the different plots of Fig. 10) as shown globally in Sect. 3 and strongly weakens the quality of the hindcast (see value of the skill score *r* in the legend of the different plots of Fig. 10). Once again, the uncertainties or the noise amplitudes, estimated as the average intra-ensemble standard deviations, reach very similar levels in the three experiments.

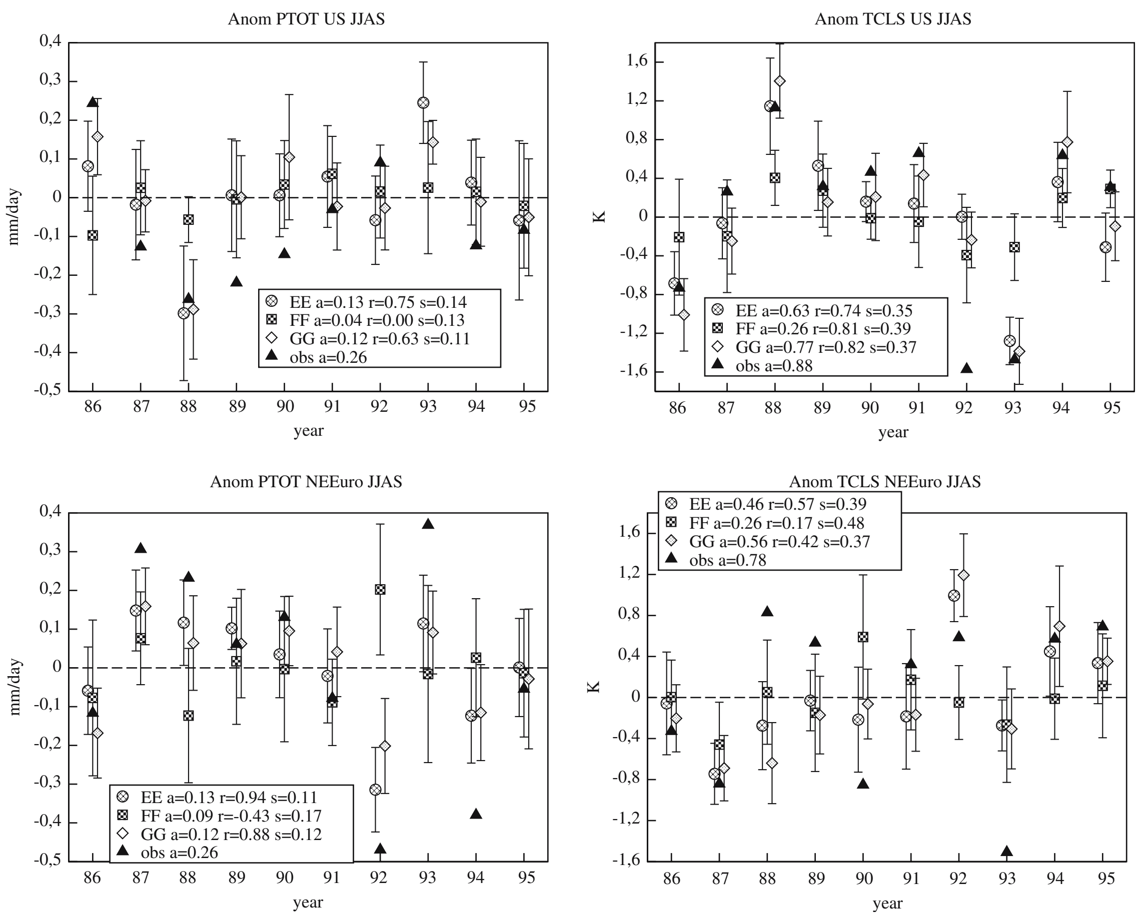


**Fig. 10** Regional average of summer (JJAS) precipitation (*left panels*) and 2 m temperature (*right panels*) anomalies. The region covering part of the USA (Eastern Europe) (see Fig. 1 for the exact definition) is shown in the *upper (lower) panels.* The *boxes* shown in Fig. 1 are defining the two regions. The ensemble mean anomalies are plotted with a *symbol* and *error bars* represent the standard deviation of the anomalies within the ensemble. In the legend the a value is the standard deviation of the ensemble means and of the reanalysis. The *r* value is the correlation of the ensemble mean anomalies with the reanalysis. The *s* value is the average standard deviation within the ensemble. Units: mm day$^{-1}$ and K

During the 10-year period covered by the GSWP2 database, 1986–1995, the North American continent has experienced contrasting warm/dry (1988) and cool/wet (1993) situations. Similarly, the European region has experienced contrasting warm/dry (1992, 1994) and cool/wet (1987, 1993) situations. In the following, we will present how well simulated those situations are in the three experiments. Our purpose here is not to discuss the processes involved in the generation and maintenance of such events, but to show the importance of the SM forcing for a proper representation of these cases.

#### *4.2.1 North American region*

The observed anomalies and the ensemble mean anomalies of the three experiments for the contrasting years 1988–1993 of precipitation, near surface temperature and 500 hPa geopotential height (Z500) are shown in Fig. 11. Large positive 2 m temperature anomalies are located over the mid-latitudes and are associated with negative precipitation anomalies of slightly smaller extent. A typical Rossby wave train with four centers is present over the North Pacific–North America sector, similar to the El Niño composite anomalies. Prescribing realistic SM forcing allows the Arpege model to create and maintain a realistic climate response with a well-positioned precipitation and temperature anomalies (SACC of 0.62 and 0.84). The forced Z500 anomalies are also well reproduced (SACC of 0.74), but are weaker than in the observations. As large parts of the precipitation and geopotential variability are chaotic (of intrinsic atmospheric dynamical origin), it is not surprising that the forced atmospheric response is weaker than the observations since the internal chaotic part has been filtered out.

The effects of the SST forcing on precipitation and temperature are weaker, maintaining small drying and warming over the central US but slightly displaced eastward. In the same vein, the large-scale dynamical response, seen by the Z500 anomalies, is a typical Rossby wave train but quite misrepresented, particularly in the eastern part of the domain. The misrepresentation of the Rossby wave type response to ENSO has also been shown in Cassou and Terray (2001) using a previous version of the same AGCM and is common to many other AGCM. This poor simulation of the major atmospheric teleconnections is a serious drawback for dynamical seasonal forecasting, particularly in winter when the teleconnections are the most active, but also in summer as revealed above, and have led to the use of statistical correction (Feddersen et al. 1999) or calibration (Doblas-Reyes et al. 2005) in the seasonal forecast community.

The combination of the SM and SST forcings is approximately linear for temperature and precipitation but is non linear for the Z500. SM forcing has a predominantly local thermodynamical and hydrological control on 2 m temperature and precipitation while SST forcing acts through modification of the large-scale atmospheric circulation. Liu et al. (1998), using a linear model, have shown that diabatic heatings play a major role in the maintenance of the large-scale circulation anomalies. On the other hand, the SM forcing through its effects on the surface fluxes and on temperature and humidity profiles is able to maintain a significant Z500 response whose amplitude is similar to the SST effect. Such a Z500 response might be explained by a change in the transient eddy activity and anomalous diabatic heating (Liu et al. 1998). Similarly the SST forcing (and surely tropical Pacific anomalies) is able to maintain a significant Z500 non-linear response (Hoerling et al. 2001).

Trenberth and Guillemot (1996) have investigated the physical processes involved in the maintenance of these two North American contrasting summers. They have revealed that the tropical Pacific SST anomalies are a major external forcing of the large-scale atmospheric circulations during these two events, and Liu et al. (1998) have particularly highlighted the predominant contribution of the western North American diabatic heating.

Our results suggest that SST forcing, and predominantly tropical Pacific SST anomalies, but also SM anomalies have certainly contributed to the set-up and maintenance of the large scale atmospheric circulations during those 2 years, as also indicated by the study of Atlas et al. (1993). Though weakened by a poor simulation of the large-scale teleconnections, our experiments suggest that, in nature, the externally forced signal is potentially a complex non-linear response to SST forcing and SM feedbacks. This forced signal is embedded in internal and chaotic variations of the atmospheric circulations, this internal variability being dependent on the initial states and not on the boundary conditions. Sud et al. (2003) have also shown evidence of the dual influence of the SM and SST factors in the generation of 1988 drought. Furthermore, they have demonstrated the necessity to correct the model systematic biases in order to track down the SST influence.

#### *4.2.2 European region*

The composite anomalies of precipitation, near surface temperature and Z500 over the European sector for the two contrasting years 1992 and 1987 are shown in Fig. 12. Similar results where obtained by compositing

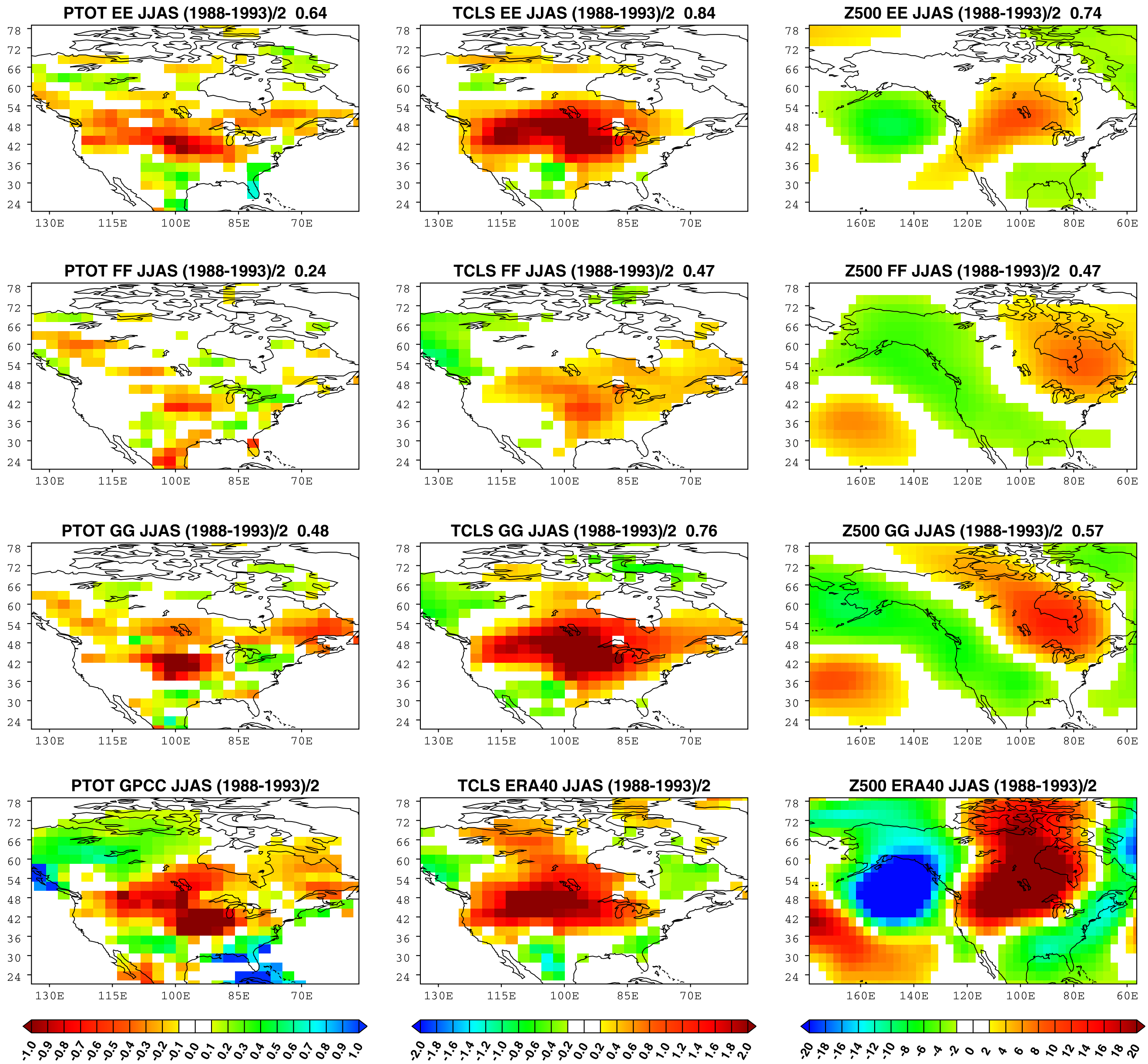


**Fig. 11** Difference between the summer 1988 and 1993 anomalies of precipitation (*left panels*) and 2 m temperature (*middle panels*) and 500 hPa geopotential height (*right panels*). The ensemble mean anomalies in the experiments are shown in the 3 *upper panels* (EE, FF and GG from top to bottom), while the observed anomalies are presented in the *lower panels*. The GPCC (precipitation) and ERA40 (2 m temperature and 500 hPa geopotential height) reanalyses are presented in the *lower panels*. The spatial anomaly correlation coefficient between the ensemble mean and reanalysis anomalies is given in the title of each plot. Units: mm day$^{-1}$, K and m

years 1992 and 1994 (warm/dry) and years 1987 and 1993 (cool/wet). A large drought is seen over the northeastern Europe in opposition with wet anomalies around the Mediterranean region and above the British islands. The drought is associated with a warm anomaly over northern Europe while the Mediterranean region and the eastern part of the domain have experienced a slight cooling.

The importance of SM in the generation of climate anomalies is also revealed over the European region. In the FF experiment where the SM is evolving freely and the SST is the only forcing, the correlations between observed and simulated anomalies are small and insignificant (except for the 2 m temperature). The atmospheric response to the SST forcing alone is not able to capture the spatial structure and the

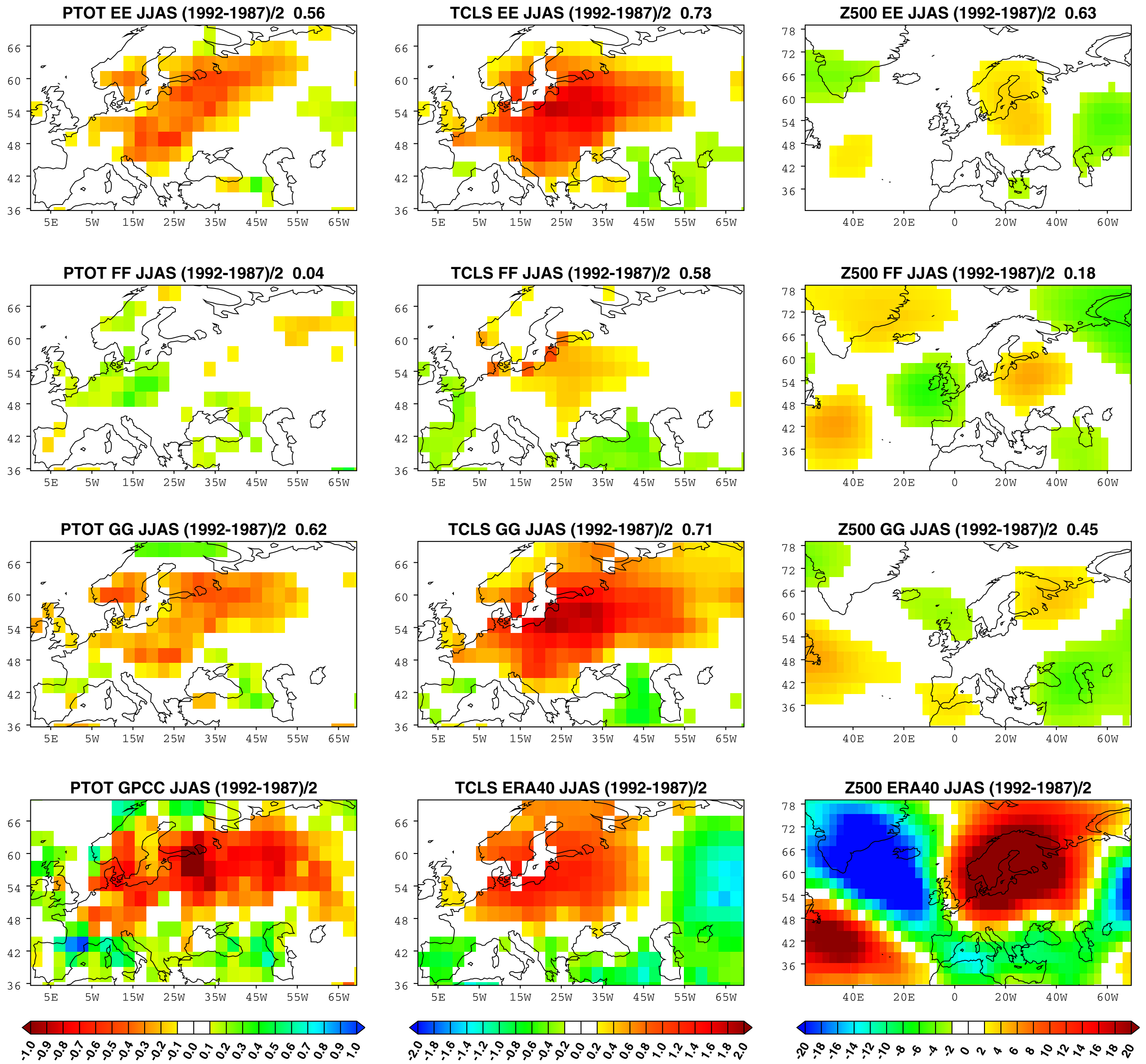


**Fig. 12** Difference between the summer 1992 and 1987 anomalies of precipitation (*left panels*) and 2 m temperature (*middle panels*) and 500 hPa geopotential height (*right panels*). The ensemble mean anomalies in the experiments are shown in the 3 upper panels, EE, FF and GG from top to bottom. while the reanalysis are presented in the *lower panels*. The GPCC (precipitation) and ERA40 (2 m temperature and 500 hPa geopotential height) reanalyses are presented in the lower panels. The spatial anomaly correlation coefficient between the ensemble mean and reanalysis anomalies is given in the title of each plot. Units: mm day$^{-1}$, K and m

magnitude of the observed temperature and precipitation anomalies. The introduction of a realistic SM forcing is leading to a better representation of the temperature and precipitation patterns. The temperature and precipitation forced responses in the GG ensemble are very close to the ones in the EE ensemble. The SACC of the temperature and precipitations anomalies are respectively 0.73 and 0.56 in EE while they are 0.71 and 0.62 (0.58 and 0.04) in GG (FF).

The Z500 anomalies maintained by the SST forcing have larger amplitude than the Z500 response to the SM. However, the spatial structure of the Z500 response to SST is quite unrealistic (SACC = 0.18). The SM forcing is able to maintain a weaker response but with a realistic spatial distribution. The combination of the two forcings is non-linear and leads to a response shifted to the NE compared with the observations.

Ferranti and Viterbo (2006) have recently evaluated the role of SM initial conditions in the severity of the European climate of 2003. They particularly assessed the relative influence of SM versus SST conditions in an idealized framework. The use of idealized SM dry perturbations helps to maintain a warmer and drier summer for up to 3 months, with a larger atmospheric response for drier situations. They also confirmed that the SST forcing had a much weaker effect than the SM forcing.

## 5 Conclusion

Three complementary GCM ensembles have been designed to assess the relative influence of SM and SST on climate variability. The GSWP2 archive has been used to create monthly SM maps over the period 1986–1995 fully consistent with the LSM (ISBA) used in the GCM simulations. While this consistency is very important given the difficulty to transfer a SM product between different land surface models, it must be emphasized that the interannual variability of the ISBA GSWP2 reanalysis is in good agreement with the multi-model GSWP2 and the ERA40 reanalyses. We have focused on two particular aspects of the variability, the potential predictability (reproducibility), and the effective predictability (hindcast skill score) evaluated against the available observations and reanalysis. The potential predictability quantifies the ratio of the forced variability (attributed to the boundary conditions) over the total variability (forced and internal variability). It can be understood as a measure of the coherency of a common forced signal in the ensemble of simulations. The effective predictability, measured as a hindcast skill score computed against the available observations, gives an estimation of the reliability of the common forced signal.

The annual cycles of potential predictability and hindcast skill have revealed that SM is contributing to a significant enhancement of the predictability mostly during the summer in the mid-latitudes. While the SST forcing enhances the potential predictability in the tropical regions, the SM forcing is the major contributor to the potential predictability in the mid-latitudes. Concerning the hindcast skill scores, the near surface temperature skills are, not surprisingly, much higher than the precipitation skills. A precipitation (2 m temperature) skill around 0.5–0.6 (0.7–0.8) is reached over the North American and Eurasian continents in response to the SM forcing. The prescribed realistic SM is essential to maintain near surface climate anomalies, at the same time reproducible (with a good signal to noise ratio) and effectively predictable (with a good hindcast score). Focusing on the North American and European regions, we have demonstrated the role of SM in the generation and maintenance of climate signals, particularly on their spatial structures, their time evolutions and their amplitudes.

In the FF ensemble, the SST is the only forcing and soil moisture is fully interacting with the atmosphere and the ensemble mean anomalies are thus an estimator of the SST forced signal. In absence of any realistic SM information, the simulated forced anomalies have weak amplitudes and their spatial structures are unrealistic. The contrasting poor (good) skill and weak (stronger) signal obtained in the FF (EE and GG) experiment suggest two main interpretations of the process involved in the maintenance of the observed anomalies. (1). The observed near-surface climate anomalies are made of a remote response to SST (mainly through advection and teleconnections) modulated by the local land–atmosphere feedback. In this case, if the model is not able to adequately simulate the atmospheric response to the SST then the interactions with the land–surface will also be misrepresented and thus the skill will be low. (2). The land–atmosphere interactions themselves are the major contributor to the observed anomalies. In this case, the initial conditions should be essential to predict the land–atmosphere state and thus the skill would probably be low in absence of any initial conditions information. It should be also remembered that even with a good knowledge of the initial conditions and a perfect land–atmosphere model, predictability of the land surface state could be low, particularly if non-linear interactions are involved.

In a different but analogous set of experiments, Kanae et al. (2006) have shown the role played by SM in the potential predictability and hindcast skill of boreal summer precipitation, mainly over semi-arid regions, where latent and sensible heat fluxes are negatively correlated. The same regions were pointed out by Koster and the GLACE team (2004b) as being "hot spots" of the land–atmosphere coupled system. Our experiments are similarly highlighting the role of SM and of land–atmosphere interactions but over a larger fraction of the mid-latitudes continents during summer, particularly over the Eurasian continent. Our results are certainly model dependent; however, as we have evaluated the hindcast skill, we suggest that, in these regions, the land surface and its interactions with the atmosphere have contributed to the generation and maintenance of climate signals over the ten summers from 1986 to 1995.

In the present study we have estimated the upper limit of predictability, arising from the use of observed (known) boundary conditions. In the dynamical seasonal prediction context, these boundary conditions cannot be anticipated and the predictability is thus dependent on the initial conditions. A particular (linear) aspect of the sensitivity of the boundary conditions on the initial conditions can be crudely evaluated as the memory or persistence of the anomalies. In a model based study, Wu and Dickinson (2004) have shown the dependence of SM memory on the season and latitudes and they confirmed most of the time-scale identified by Manabe and Delworth (1990). In situ SM observations also indicate typical SM anomalies time-scales of 2–3 months in the northern mid-latitudes but much longer during winter (Entin et al. 2000). A secondary maximum in the SM memory was also evidenced in the previous studies. Schlosser and Milly (2002) using a different model concluded that, in the northern mid-latitudes, while SM memory and predictability is stronger in winter, the potential for atmospheric predictability is the largest during the summer months. Koster and Suarez (2003a) have revealed that the summer precipitation anomalies are conditioned, at least to some extent, by the 2 month earlier anomalies. Furthermore this conditioning can be partly reproduced by atmospheric model if and only if land–surface interactions are active.

In order to assess the relative importance of SST and SM initial conditions, the SM and SST memory can be compared using the ISBA and multi model GSWP2 reanalyses and the HadSST climatology. The summer SM and SST persistence, computed as the 2 months autocorrelation of monthly anomalies over the period 1986–1995 (very similar summer SST persistence was obtained by using the 1950–2000 period), are presented in Fig. 13. This figure shows how well correlated are the summer (JJAS) SM and SST monthly anomalies with the former AMJJ SM anomalies (2 months before). Significant lagged correlations of SM anomalies are reached over large parts of the northern midlatitudes, meaning that part of the initial anomalies are persisting in the next 2 months. The similarity of the ISBA and multi-model SM persistence is particularly strong over the Eurasian and African continents, suggesting that the SM memory in ISBA is robust over those regions. Nevertheless, some differences are noticeable over the American continents, the ISBA SM persistence being stronger over the Amazonian forest and weaker over the US Great Plains.

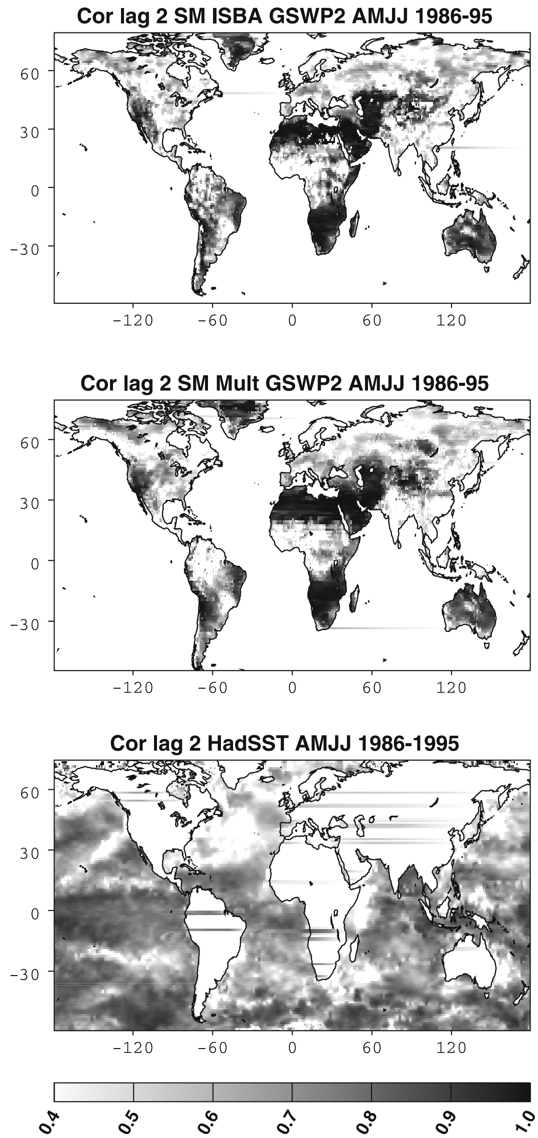


**Fig. 13** Two-month lagged auto-correlations of SM anomalies in the ISBA (*upper panel*), multi model (*middle panel*) GSWP2 reanalyses and of the SST anomalies in the HadSST data (*lower panel*) for the AMJJ season. This corresponds to the persistence of the SM and SST anomalies between the AMJJ season and the JJAS season

Contrasting with most of the tropical and southern oceans, where summer SST anomalies persistence is high (2 months lagged autocorrelation over 0.7), the North Atlantic and the northwestern Pacific oceans, surrounding most of the northern midlatitudes continents, show quite low (below 0.5) SST persistence. While SM persistence is quite heterogeneous in space, some regional patches of high persistence are noteworthy over the northern Hemisphere summer midlatitudes. The contrast between the smaller persistence of SST and the longer persistence of SM anomalies over some northern midlatitudes regions is suggesting that the forecast, or more specifically the hindcast, of summer climate anomalies may benefit from the use of realistic initial SM conditions.

It seems thus critical to assess what could be the prediction scores attained with realistic land/ocean initial conditions, and more specifically what is the predictability of the SM anomalies themselves. Since the impacts of SM seem to be mostly local, a comparison of Fig. 7 (showing the actual skill of the AGCM in hindcast mode, e.g., with known boundary conditions such as SM) and Fig. 12 (showing the SM memory) may reveal a first gross estimation of the skill that we can expect in forecast simulations using realistic SM initial conditions. But a new set of GCM experiments will be designed to address the initial conditions problem and particularly the role of realistic SM initial conditions on the boreal summer predictability. The recently available GSWP2 dataset offers the opportunity to evaluate the impacts of realistic spring and early summer SM anomalies on the subsequent summer climate.

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