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Summary

The impact of Atlantic multi-decadal variability on European climate has been investigated in a 500-yr long coupled model simulation and ensemble uncoupled atmospheric model simulations for the 20th century.

First, the impact of multidecadal variability of the North Atlantic thermohaline circulation (THC) on the European climate has been analyzed using a 500-years control simulation with the global coupled atmosphere-ocean general circulation model ECHAM5/MPI-OM, which realistically reproduces THC variations. The strongest correlations with precipitation, sea level pressure (SLP), and surface air temperature are found for boreal winter season, whilst summer season in general does not show significant correlations. A positive phase of the THC is related to the increase of precipitation (decrease of SLP) over Norwegian-Barents Seas and northern Eurasia and warming over North Atlantic, Europe and northern Asia. Intensified advection from the ocean to the western part of Eurasia is suggested to be responsible for these changes. Another mechanism is an increase of the oceanic inflow to the marginal Arctic seas with corresponding sea ice retreat. This causes particular strong correlations in the Norwegian and Barents Sea areas. Changes in the sea surface temperature associated with the THC can affect the large-scale atmospheric circulation in the Northern Atlantic. Model results show significant modulation of the Iceland Low intensity by the North Atlantic THC on multi-decadal timescales.

It is shown using observations and the ECHAM5/MPI-OM simulation that a correlation between the North Atlantic Oscillation and temperature variations in the northern high latitudes (north of 60°N) is not stable, and may vary from statistically significant positive to negative values. According to model results, the Arctic climate exhibits a robust connection to the Barents Sea inflow (BSI), which may explain a major part of decadal variability in the Arctic. The interannual variability of the BSI is determined by local atmospheric circulation, whilst multi-decadal changes are associated with the corresponding variations on the THC. The latter may be responsible for the strong multi-decadal natural climate variations in the Arctic and northern Europe, such as the early 20th century warming anomaly and, possibly, for substantial part of the ongoing warming during the last decades.
Second, the influences of anthropogenic climate change versus internal natural climate variability on summer night and day time temperature over France was investigated. Using a high-quality dataset of observed temperature and large ensembles of high-resolution atmospheric model simulations, a significant human influence on the 20th century evolution of summer night and day time temperature over France is detected. In addition to the anthropogenic-induced changes, the regional analysis indicates that natural oceanic forcing associated with Atlantic Multidecadal Variability (AMV) contributed to the accelerated warming in France during the last decades.

**Introduction**

The North Atlantic thermohaline circulation (THC), a part of a global 3-dimenional belt of ocean currents [Manabe and Stouffer 1999], is responsible for about 50% of the meridional poleward oceanic heat transport in the Northern Hemisphere and is partly responsible for the relatively warm European climate. Variations of the THC manifest themselves in significant oceanic SST anomalies in the northern Atlantic [Latif et al. 2004] and may modulate the warm Atlantic currents entering Arctic. This implies a significant potential importance of the THC variability for the climate in the European region and in the Arctic.

According to the model estimates, anthropogenic global warming can lead to significant variations in the intensity of the THC [IPCC 2007]. Possible variations in the THC and their regional consequences are important issues in climate research. An increase in the surface temperature and precipitation in the Arctic region would decrease salinity and density in the surface layer of the Arctic Ocean and Arctic seas. This tendency is enhanced by melting of sea ice and by increased river discharge into the Arctic basin. As a result, the stability of the upper oceanic layer should increase, damping high-latitude oceanic convection and decreasing intensity of the THC. This is accompanied with a large-scale cooling anomaly in the northern North Atlantic, which may mitigate anthropogenic warming in Europe. Observed SST variations during the last 130 years indicate the THC is also characterized by strong multidecadal natural fluctuations. These SST variations are of order 1°C, and cover the whole northern North Atlantic [Latif et al. 2004] and may be predictable on decadal time scale [Keenlyside et
al. 2008]. Observational hydrographic data on the strength of the THC exist only for the last two decades [Lumpkin and Speer 2003].

A close relation between the THC and SST [Latif et al. 2004] makes it possible to evaluate the THC variability back to the middle of the 18th century. However, even with such a time-series it is difficult to obtain statistically robust estimates of the THC-climate teleconnections. To do this two approaches are followed here: First, a simulation with a coupled climate model, which realistically reproduce THC and North Atlantic SST variability, are used. Second, a detection-attribution approach is applied using high-resolution atmospheric model simulations forced with observed SST, sea-ice cover, and evolution of greenhouse gas and sulphate aerosol concentrations, as well as other natural forcing factors. Full details of this second study can be found in Terray and Phlanton (2008).

A link between multidecadal variations of the North Atlantic THC and European climate in a control simulation with the ECHAM5/MPI-OM model

Here, monthly mean SST, sea level pressure (SLP), and precipitation (R) from the 500 yr control (without external forcing) simulation with the coupled atmosphere-sea ice-ocean general circulation model (GCM) ECHAM5/MPI-OM are analyzed. Concentration of greenhouse gases and sulfate aerosols in the atmosphere were constant, at levels corresponding to the present-day climate. The intensity of the THC was characterized by the annual mean maximum meridional overturning (MOC) in the North Atlantic at 30N. The amplitude and periodicity of the multidecadal variability in the North Atlantic in this simulation fits very well to observations [Latif et al. 2004], and the model resolves complex feedbacks between the THC and Arctic heat and fresh water balance [Jungclaus et al. 2005].

The atmospheric model ECHAM5 [Roeckner et al. 2003] has T42 spectral resolution (approximately 2.8°) and 19 vertical levels (up to 10 mbar). The oceanic model MPI-OM [Marsland et al. 2003] has variable spatial resolution (from 10 km near Greenland to 300 km in the Pacific Ocean) and 40 vertical layers. A detailed description of the model and control experiment is given in Jungclaus et al. [2005]. High spatial resolution in the North Atlantic facilitates a better representation of deep convection in
the Greenland Sea. The results of the last 450 yr of the model experiment after stabilization of the mean climate in the model are analyzed.

Correlations between variations in the annual mean MOC index and the SST, SLP, and R were computed for different seasons. Running average values with different intervals of averaging were used to analyze low-frequency variations. The strongest correlations were found in the winter period. This is related to the most intense ocean surface–atmosphere and ocean–land heat fluxes in the cold period of the year owing to the strongest zonal circulation in the troposphere in the extratropics. Figure 1 shows the spatial distributions of correlation coefficients (r) between low-frequency variations (20 yr running means) of annual mean MOC index and seasonal values of SST (Fig. 1a), SLP (Fig. 1b), and R (Fig. 1c) for winter (December–February). The absolute correlation values exceeding 0.45 in Fig. 1 are statistical significant at 5% level.

The regions of positive correlation between winter surface temperature and annual mean intensity of the THC prevail in the Northern Hemisphere. Figure 1a shows elongated regions with a high positive correlation between MOC and the winter surface temperature (SAT) over the North Atlantic from the tropical region to the arctic latitudes, over Europe (and south of Europe), and in mid-latitudes in Asia. A negative correlation between the sea level pressure and MOC dominates in the Northern Hemisphere, including middle and high latitudes over Northern Eurasia and high latitudes over the North Atlantic and Arctic Ocean (over the Norwegian and Barents Seas) (Fig. 1b). High correlation is found, in particular, in the area of the Siberian anticyclonic High and Iceland cyclonic Low.

Positive correlation between winter precipitations with the MOC is found over Eurasia and North Atlantic (Fig. 1c). According to Fig. 1, the region of significant correlation between the intensity of the THC and precipitation generally corresponds to the region of the highest correlation between the intensity of the THC and SLP with some relatively small-scale features (Fig. 1c). It is worth noting that despite the vicinity of the SST anomalies in the North Atlantic, no significant correlation is found in Western Europe for SLP and R. However, the high correlation region spreads from the Norwegian and Barents seas to the East Siberia.

According to Fig. 2, a significant correlation (0.7) is found between MOC and the intensity of the Iceland Low, characterized by stationary negative anomaly of the
SLP center close to Iceland. An intensification (deepening) of the Iceland Low during intensification of the THC is related to the displacement of the trajectories of the North Atlantic cyclones and their intensification. This was found from the NCEP/NCAR reanalysis data in the last 40 yr of the 20th century [Gulev et al. 2001], which was a period with a positive SST trend in the North Atlantic.

It is worth noting that the correlation with the intensity of the THC for other seasons (in particular, in summer) is lower than in winter. It is important to note that the strongest variations are prominent in winter in recent decades (based on observations). In winter, when the meridional differences between temperatures at low and high latitudes are greater and the tropospheric circulation is more intense (including those with strong zonal transport at mid-latitudes), larger scale variations in the climatic variables (including temperature anomalies) are better manifested. When the THC intensifies, the temperature in the North Atlantic increases, and heat anomalies spreads more effectively in winter to European and Asian regions due to the zonal transport at mid-latitudes. A pressure decrease in the regions of the Iceland Low and Siberian High facilitates the manifestation of such a tendency. This corresponds to intensification of the cyclonic circulation at subpolar latitudes in the North Atlantic. An increase in the pressure gradient between lower and higher latitudes should lead to intensification of the geostrophic wind, i.e., intensification of the heat-and-moisture transport from the Atlantic to Europe with positive anomalies of temperature and precipitation. Of course, the spatial structure of the anomalies of the precipitation field is much more variable than that of the temperature field. This is reflected in specific features of its correlation with the variations in the THC (Fig. 1). Weakening of the winter Siberian High favors the deep penetration of this influence into the continent and even to the eastern coast of Asia. Remote influence of the THC on the regions of East Siberia can also be related to the variations in the ice cover of the Arctic Ocean (in particular, the Barents Sea) caused by variations in the oceanic heat influx from the Atlantic [Bengtsson et al. 2004]. Variations in sea-ice cover area modulate intense fluxes of heat and moisture from the sea surface to the atmosphere. Then, the fluxes are transported in the southeastern direction, resulting in the formation of the anomalies of surface temperature and pressure, in particular, in the region of Siberian High.
Impact of the THC on multidecadal climate variability in the northern high latitudes in the ECHAM5/MPI-OM control simulation

As seen in Fig. 3, the ECHAM5/MPI-OM coupled model adequately reproduces the SAT variability in the Arctic region (60–90N) during the winter period (November–April). Model anomalies of the SAT have the same amplitude as the observed values, and their characteristic time scale is also 70–80 yr. The analysis given in this work is related to the values averaged over the winter period, because winter anomalies are two to three times larger than the summer ones and they make the main contribution to the mean annual values of the anomalies. Variations in the SAT are closely related to the variations in the sea ice cover (SIC) (correlation coefficient is –0.77 for instantaneous values and –0.89 for time series smoothed with a 11-yr moving averaging). The map of regression coefficients of the SAT anomalies over the anomalies of the Arctic SIC in the model (not shown) demonstrates the maximal values over the Barents Sea. The Barents Sea is characterized by significant heat losses from the ocean to the atmosphere (on average, approximately 200 W/m² for the ice-free sea surface), which are modulated by variations in the SIC. The anomalies of the SAT in the Barents Sea region also made the largest contribution to the long-term Arctic temperature variability during the whole of the 20th century, presumably caused by fluctuations in the SIC [Bengtsson et al. 2004; Semenov and Bengtsson 2003].

The last three decades of the 20th century were marked by the highest increase in the North Atlantic Oscillation (NAO) (over the instrumental observation period), which is the leading mode of the atmospheric circulation variability in the Atlantic sector of the NH [Wanner et al. 2001]. A transition to the positive phase of the NAO explains a significant part of climatic variations in Europe and in the NH as a whole. It was supposed that this intensification was responsible for the simultaneous warming in the Arctic region [Moriz et al. 2001]. However, the long-term NAO variability cannot explain the Arctic warming in the mid 20th century [Bengtsson et al. 2004]. As was shown in [Osborn et al. 1999], the correlation between the NAO and SAT in the extratropical latitudes is not stationary and changed from significant positive correlations to zero. Figure 4 shows the correlation between NAO and temperature in the Arctic region during the winter period with a running 50-yr window. It is seen that the significant positive correlation between the NAO and SAT anomalies was observed only at the end of the 19th century and in the second half of the 20th century and
changed to –0.2 in the middle of the 20th century. The ECHAM5/MPI-OM model adequately reproduces the unstable character of the link between the NAO and Arctic temperature, demonstrating similar secular oscillations of the running correlation even with the possibility of statistically significant negative values. As was suggested in [Bengtsson et al. 2004], the variations in the oceanic inflow to the Barents Sea could cause long-term variations of the SAT in the Arctic region. This is also suggested by the analysis of the temperature anomalies at high latitudes of the NH [Semenov and Bengtsson 2003] that demonstrates the strongest anomalies in the SAT related to long-term variability in the Barents Sea region. The maximum of the winter Arctic SAT variability in the model is also located in this region.

The inflow of relatively warm Atlantic water into the Arctic Basin results in the corresponding variations in the SIC. The model adequately reproduces the oceanic inflow into the Barents Sea (the volume flux across 20E) equal to 2.1 Sv, which falls in the range of the empirical estimates [Simonsen and Haugan 1996]. The root-mean-square deviation of the SIC in the Barents Sea in winter in the model is 0.12x10^6 km^2 or 15% of the climatic mean value. This value accounts for more than 40% of the total SIC variability in the Arctic and Atlantic basins. The correlation between the oceanic inflow and SIC in the Barents Sea over the 450-yr period of the model experiment is equal to –0.78 (–0.91 for the 11-yr moving averaging) and –0.55 (–0.79) for the SIC in the Arctic and Atlantic basins. This indicates the determining role of the oceanic inflow in the variations of the SIC, which causes the SAT variations in the winter period. This correlation also varies, but it is much more stable than the correlation between the NAO and Arctic temperature considered above. Figure 5 presents the values of the running correlation in an 80-yr window between the time series of the NAO, Arctic SAT, and oceanic inflow into the Barents Sea in the model (for time series initially smoothed with 5-yr running mean). It is seen that the correlation between the SAT and oceanic inflow is positive and statistically significant practically over the entire experiment period (excluding a 50-yr time interval). At the same time, the correlations between the SAT and NAO are strongly variable and their variations are similar to the corresponding variations in the running correlation between the NAO and oceanic inflow. For example, in the model years 2100–2200, the Arctic SAT–NAO and NAO–oceanic inflow lack virtually any correlation (Fig. 5), while the correlation between the inflow and SAT reaches 0.8. In general, the Arctic SAT and NAO are correlated during the
same periods when a significant correlation between the NAO and the inflow to the Barents Sea occurs.

Thus, the data of the model experiment suggest that the variations in the oceanic inflow into the Barents Sea play a crucial role in the formation of the SAT variability (on a decadal and longer time scale) in the Arctic region during the winter season. The intensity of the oceanic inflow determines the SIC that modulates turbulent heat fluxes from the sea surface to the atmosphere. According to the observational data and model experiments (see [Bengtsson et al. 2004] and references therein), the variations in the atmospheric circulation in the region of the Norwegian and Barents seas are the main factor determining the interannual variability of the oceanic inflow. The gradient of the atmospheric at sea level pressure between northern Norway and Spitsbergen is closely related to the oceanic inflow.

For longer time scales (a decade and more), the variations in the North Atlantic thermohaline circulation (THC) also play an important role. The contribution of the THC becomes crucial for long-term (secular) oscillations. In the model experiment described here, the correlation between the inflow to the Barents Sea and the index of the THC (maximum of the meridional overturning in the North Atlantic at 30N) is equal to 0.28 for the annual mean values and 0.60 for the 30-yr running averaging (for the duration of the time series equal to 450 yr). The corresponding values for the correlations with the atmospheric pressure gradient are equal to 0.42 and 0.29. The co-variability of low frequency THC fluctuations and Arctic climate is illustrated by the Figure 6, where low-pass filtered time series of the annual THC index, Arctic temperature and Barents Sea inflow are shown.

The variations in the ice cover in the Barents Sea strongly influence the regional atmospheric circulation, which, in turn, can influence the intensity of the Arctic anticyclone and ice transport from the Arctic region to the Atlantic. This fact supports the possibility of feedback between the regional climatic variations in the Barents Sea and North Atlantic THC.

The influence of Atlantic Multi-decadal variability on France summer temperature

Wavelet analysis of observed annual mean air temperature over France shows that the 20th century variability is dominated by an almost monotonous secular trend modulated

The origins and spatial signature of recent decadal summer temperature changes in France are investigated, using the standard “frequentist” approach for the detection and attribution [IDAG 2005]. Furthermore, we also rely on the forced atmospheric general circulation model (AGCM) detection framework [Sexton et al. 1999] in which the definition of signal and noise differ from the standard one. We first carry out the detection step by showing that observed changes are significantly different than can be explained by natural atmospheric internal variability. We then quantify the relative contributions of the various external (natural and anthropogenic) forcing factors to observed changes (attribution step) and suggest the possible physical processes responsible for the space-time climate change sub-regional pattern in both observed and simulated data (physics coherency step).

Here, we use the ARPEGE AGCM [Gibelin and Déqué 2003] with high resolution (60 km) over Europe. The model is forced with monthly mean observed sea surface temperature. The observed greenhouse gas and sulphate aerosol concentrations as well as natural forcing factors. For a given ensemble, the ensemble members (six for each ensemble) have the same forcing and different atmospheric initial conditions. The various ensembles are all forced with the same observed changes in sea surface temperature and sea-ice extent over the 1950-1999 period, but differ in terms of the included combinations of natural and anthropogenic effects. The SST ensemble has anthropogenic and natural forcing factors fixed at their 1950 value. The other three ensembles we consider here, G, GS and GS-NAT, are made by successively including the observed evolution of greenhouse gases (GHG), sulphate aerosols (SUL) concentrations and changes in natural forcing factors including both solar irradiance (SOL) and stratospheric aerosols following volcanic eruptions (VOL). Here and thereafter, we refer to signals G, GS, GS-NAT as the combined effect of (a) SST and GHG (b) SST, GHG and SUL and (c) SST, GHG, SUL, SOL and VOL, respectively.

The history of temperature changes averaged over France is well reproduced by simulations of the ARPEGE model, when driven by SST and anthropogenic forcing (Fig. 7). No significant difference is seen between the SST (G) and GS ensemble means for France averaged temperature apart from the last decades, where the GS signal seems
to slightly emerge. The GS ensemble mean 30-year (1970-1999) linear trends have significantly larger amplitude than their SST counterpart, in particular for maximum temperature (0.3 versus 0.04 K/decade) which suggests a possible influence of anthropogenic forcing on the recent warming (Fig. 8). Both ensembles (G & GS) show a clear simulated warming for minimum temperature (with the GS value being 50% stronger) while the SST ensemble shows weakly positive or even slightly negative trends for maximum temperature, thus contrasting with the GS ensemble. A qualitative estimate of the range of the France warming rate due to the GHG forcing can be made. If one assumes no anthropogenic origin for the oceanic signal, the GHG forcing is only due to direct radiative forcing. The related warming rate can then be estimated by the G-SST difference in temperature (simply estimated from the sum of Tmin and Tmax) trends. The result is 0.13 K/decade and can be viewed as a lower bound of the GHG forcing effect. An upper bound might also be derived by assuming that the oceanic signal is entirely of anthropogenic origin. The GHG forcing effect can then be estimated by the G ensemble mean temperature trend which is 0.26 K/decade. The above GHG-related warming rate range is coherent with the global land observational estimates derived by Wild et al. (2007).

A detection-attribution analysis is applied to all possible signal combinations in order to find which ones are consistent with the observed record over the various periods. Combinations that fail at the 10% level are rejected. In addition, we also reject those signal combinations in which another signal, not present in the combination, is detected when added to the tested combination (Stott et al. 2001). The first result is that the SST signal on its own is inconsistent with the observations for both Tmin and Tmax. Internal atmospheric variability and oceanic forcing alone cannot thus explain the evolution of France summer temperature during 1950-1999. The second result with regard to Tmin, is that we have to reject combinations which do not include SUL, for instance G and (SST, G), as GS is detected when added to the latter. In the case of Tmax, the robustness of the SUL influence is much less clear (SUL is only detected in one case, within the (G, GS) combination, and the lower uncertainty bound is close to 0) in contrast with that of GHG. The influence of direct anthropogenic forcing factors (GHG and SUL) is thus detectable in addition to that of SST and has to be invoked to explain the late century minimum and maximum temperature warming.
Regarding the oceanic contribution to the low frequency variability, we suggest that one of the inter-decadal components is linked to AMV while the other reflects the regional oceanic signature of climate change and in particular the strong warming over the east Atlantic after 1980. Can we roughly estimate the fraction of the SST-forced temperature change due to anthropogenic factors? This amounts to determine the influence of anthropogenic forcing upon the oceanic variations responsible for the summer temperature warming over France. To assess this influence, we perform a principal component analysis of the SST ensemble mean summer minimum temperature over France. The spatial pattern of the first mode depicts a warming pattern with larger amplitude over western and southern France. The mode time series can be viewed as an estimate of the time history of the SST forced component of the minimum temperature variability. The mode time series closely follows AMV up to the mid-eighties and rises more sharply afterwards (Fig. 9) suggesting a different origin for the most recent warming trend.

To partly understand the oceanic modes involved, the associated spatial pattern of sea surface temperature for the different periods are estimated by correlating the 1st mode time series with the model ocean surface temperature (which are the observed SSTs). The correlation patterns are very different patterns for the 1950-1979 and 1970-1999 periods (Fig. 10a,b). The pattern for the former period is very similar over the oceans to the one obtained with the AMV index [Trenberth and Shea 2006] (Fig. 10c). Note also that significant correlations are also observed over the Sahel and India when the index is correlated with the SST-ensemble mean surface temperature and precipitation. This is in agreement with previous studies on AMV-related teleconnections [Folland et al. 1986, Zhang and Delworth 2006]. Finally, there exists evidence from long coupled model simulations of the internal nature of the AMV related to fluctuations of the thermohaline circulation (Knight et al. 2005, and as describe above). In contrast, the correlation pattern of the latter period shows close resemblance with the global mean ocean surface temperature index (Fig. 10d). This suggests that the oceanic influence upon summer France temperature is mainly due to internal oceanic forcing for the earlier period while the latter one is dominated by the influence of external forcing through the associated global mean ocean warming and in particular its regional signature over the northeast Atlantic. It has been recently suggested that a significant fraction of the recent increase in North Atlantic temperature
can be associated to global warming [Trenberth and Shea 2006, Pierce et al. 2006]. This suggests that the low-frequency part of the detected oceanic signal is mainly due to natural internal variability linked to the AMV while the global SST increase due to human activities is the dominant term for the last two decades of the 20th century. The observed tendency towards warm SST anomalies in the near eastern Atlantic for the last two decades favours a spatially coherent warming over France which is subsequently amplified by direct anthropogenic forcing and locally modulated through a positive feedback between evapotranspiration and soil moisture changes [Terray and Planton 2008].

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References


Stott, P. A., et al., Attribution of twentieth century temperature change to natural and


Fig. 1. Correlation coefficients between multidecadal variations (20-yr running means) of the annual mean THC intensity and winter (December–February) values of (a) surface temperatures; (b) sea level pressure, and (c) precipitation.
Fig. 2. Intensity (negative of the SLP anomalies) of the wintertime Iceland Low, IL (hPa), as a function of annual mean MOC intensity (Sv), 20-yr running averaging.

Fig. 3. Anomalies of surface air temperature (SAT) in the Arctic (60°–90°N) during the winter half-year (November–April) based on (red) empirical data [Jones et al.] and (black) the model simulation (11-yr running mean). The model data are shifted in time to illustrate the similarity with the observed long-term SAT variability.
Fig. 4. Correlation between the NAO index and Arctic SAT during the winter period based (red) on observations and (black) model calculations. Correlations are computed with a running 50-yr window, after application of a 5-yr running mean to the time series. Model data are shifted in time. Dashed line (at 0.55) is for 10% significance level.

Fig. 5. Correlation (in running 80-yr window; initial time series are smoothed by 5-yr running averaging) between (1) the oceanic inflow to the Barents Sea and Arctic SAT, (2) between the inflow and NAO index, and (3) between the NAO index and Arctic SAT. Dashed line (at 0.48) indicates the 10% significance level.
Figure 6. Simulated annual mean Barents Sea inflow (black), Arctic SAT anomalies (red), and MOC index, maximal overturning at 30N, shifted to fit to the right y-axis for the Barents Sea inflow (blue), 21 year running means.

Figure 7: Time evolution of France averaged summer maximum (dashed line) and minimum (solid line) temperature anomalies over the 1950-1999 period. The anomalies are computed using as a reference period the 1961-1999 climatology. The black lines show the observed values while the red and blue lines depict the SST and GS ensemble means, respectively. For minimum temperature, the pink and light blue shading gives the members envelope (minimum and maximum values) of the GS and SST ensembles, respectively. The dots show the 1990s minimum temperature decadal means (colour shading as above).
Figure 8: Simulated linear trends over the 1970-1999 period for minimum (left panel) and maximum (right panel) temperature (a), (b) GS ensemble mean (c), (d) SST ensemble mean. The Ave inset shows the France mean trend value while Sdev indicates the standard deviation of the grid-point trend values. In (a), (b), hollow circles indicates grid points where the null hypothesis of zero trend for the GS—SST temperature difference cannot be rejected at the 5% significance level using a simple t-test.
Figure 9: Annual SST anomalies averaged over the North Atlantic (0 to 60°N, 0 to 80°W) for the 20th century, relative to 1900 to 1970 (in K) with the global mean removed (thin black line with fill). The global mean SST (averaged over the global oceans, 60°S to 60°N) is given by the heavy black line. The heavy coloured lines show the first mode time series of a principal component analysis applied to the France summer Tmin for the SST, G and GS ensemble means. All time series are first filtered with a low-pass symmetric filter with 13 weights \([1/576(1., 6., 19., 42., 71., 96., 106., 96., 71., 42., 19., 6., 1.])\) and a half-power point at 16-year periods and then standardised.
Figure 10: Correlation of various low-pass filtered indexes with annual observed SSTs based on summer (JJA) values. The observed SST dataset is the one used as boundary conditions for all the AGCM forced experiments. The applied filter is the same as in figure 9. a, b) The index is the 1st mode time series of an EOF analysis of the summer minimum temperature SST-ensemble mean for the 1950-1999 period. The a) and b) correlation maps are estimated for the 1950-1979 and 1970-1999 periods, respectively. c) The index is estimated as observed summer SST anomalies averaged over the North Atlantic (0 to 60°N, 0 to 80°W) for 1900-1999, relative to 1901-1970, but with the global mean removed. d) The index is estimated by observed summer SST anomalies averaged over the global oceans (60°S to 60°N) for 1900-1999 relative to 1901-1970.