An assessment of global and regional climate change based on the EH5OM climate model ensemble

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Abstract An analysis of climate change for global domain and for the European/ Mediterranean region between the two periods, 1961–1990 (representing the twentieth century or "present" climate) and 2041–2070 (representing future climate), from the three-member ensemble of the EH5OM climate model under the IPCC A2 scenario was performed. Ensemble averages for winter and summer seasons were considered, but also intra-ensemble variations and the change of interannual variability between the two periods. First, model systematic errors are assessed because they could be closely related to uncertainties in climate change. A strengthening of westerlies (zonalization) over the northern Europe is associated with an erroneous increase in MSLP over the southern Europe. This increase in MSLP is related to a (partial) suppression of summer convective precipitation. Global warming in future climate is relatively uniform in the upper troposphere and it is associated with a 10% wind increase in the subtropical jet cores. However, spatial irregularities in the low-level temperature signal single out some regions as particularly sensitive to climate change. For Europe, the largest near-surface temperature increase in winter is found over its north-eastern part (more than 3°C), and the largest summer warming (over 3.5°C) is over south Europe. For south Europe, the increase in temperature averages is almost an order of magnitude larger than the increase in interannual variability. The magnitude of the warming is larger than the model systematic error, and the spread among the three model realisations is much smaller than the magnitude of climate change. This further supports the significance of estimated future temperature change. However, this is not the case for precipitation, implying therefore larger uncertainties for precipitation than for temperature in future climate projections.

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1 Introduction

There are no elements of doubt that the global climate change is an ongoing process. In the period 1906–2005, the globally averaged surface temperature has changed by $+0.74^{\circ}C \pm 0.18^{\circ}C$, with almost doubled rate of warming over the last 50 years of the same period (Intergovernmental Panel on Climate Change, IPCC 2007). In a very cautious wording and with observational and forcing uncertainties taken into account, the IPCC (2007) report attributes this warming mainly to anthropogenic causes, in particular to greenhouse gas forcing. The warming is observed in both the atmosphere and the ocean when the natural external forcing factors would otherwise have produced cooling.

Coupled atmosphere-ocean general circulation models (AOGCMs), hereafter referred to as climate models, are indispensable tools for studying and better understanding of climate change and projections of future climate. The reliability of climate models has increased considerably in the past years. The confidence that climate models can provide realistic estimates of future climate is based on their ability to reproduce correctly the present climate. This was possible by improving the models in various ways—by improving dynamics, horizontal and vertical resolutions, physical parameterisation and inclusions of more processes, like aerosols, sea ice, and land surface. A large body of literature is available documenting various improvements in climate models (see for example, Randall et al. 2007).

However, despite improvements, climate models are approximations of real atmosphere, oceans, and other elements that make climate system, and as such are not free from errors. For example, Lambert and Boer (2001) and Covey et al. (2003) compared a number of AOGCMs with observation based estimates of various climate variables and found systematic differences, some of them common to many models. This implies that there is a scope for future improvements of climate models.

Projections of climate change are based on time-evolving concentration of greenhouse gases and aerosols. Since no unique development of these concentrations could be possibly foreseen, various emission scenarios are developed (Nakićenović et al. 2000). Scenarios include assumptions about e.g. growth of the world population, economic growth, energy production and consumption, which all affect greenhouse gas emissions and aerosol concentration.

An analysis of climate change based on a single global climate model may seem a drawback when compared to the multi-model approach. A rationale for the choice of one, in our case the EH5OM climate model (Roeckner et al. 2003), should be viewed in the context of the following discussion. The multi-model procedure is normally based on consistency among model results. The latest generation of climate models generally tend to be in good agreement, as found by Reichler and Kim (2008) who compared the results from three generations of the Coupled Model Intercomparison Project (CMIP) models. Whereas the multi-model mean for the CMIP3 models indicates an overall improvement when compared to individual models, this improvement is much smaller than the one seen for the CMIP2 and CMIP1 models. On the other hand, Wang (2005) pointed out that the consistency among climate models does not necessarily imply improved reliability, in particular for variables manipulated through parameterisation schemes. The response of an individual model may differ from the multi-model response because it includes an

improved parameterisation or because it includes a mechanism or feedback that the other models do not (Wang 2005).

Furthermore, Rind (2008) doubts that uncertainties in climate model responses, which are largely due to different model physics, will be reduced by a multi-model approach. Different physical parameterisations imply solving different equations and "averaging different formulations" may not produce an improved result. This is in contrast with weather forecasting where such an approach yields better results (e.g. Krishnamurti et al. 2000), because according to Rind (2008), the model response here is dominated by atmospheric dynamics. Thus, errors in numerical weather prediction models, that are solving the same dynamical equations, are bound to be minimised when averaged.

Following Reichler and Kim (2008), EH5OM model is one of the most successful CMIP3 models (out of 22) in simulating the climate of the twentieth century models. The ranking was based on the metric calculated from climatologies of 14 variables over the 20-year period. In addition, the EH5OM data provide initial and lateral boundary conditions for dynamical downscaling by a regional climate model, the programme currently underway at the Croatian Meteorological and Hydrological Service (DHMZ; Branković et al., in preparation). Therefore, to better understand climate change on finer spatial scales, it is essential to analyse both global and regional climate change as defined in a global climate model.

Although climate modelling with complex AOGCMs demands relatively large computing resources, there are examples of climate models ran in ensemble mode. In this study, we assess the three-member ensemble made by the EH5OM climate model under the IPCC A2 scenario. In terms of statistical significance and of climate predictability, we are confident that the ensemble approach applied in this paper outweighs a possible shortcoming of using the data from a single climate model, no matter how small our ensemble is. Because a single integration of a climate model could be sometime misleading (e.g. Barnett et al. 2000), ensemble (and seasonal) averaging reduces the effect of noise, which is due to internal model variability, and highlights the climate signal.

The main objective of this work is to assess the change between the "present" climate (second half of the twentieth century) and the future climate (mid twenty-first century) over a limited domain covering southern Europe, together with the changes in variability of the two climates, from the most recent version of the EH5OM AOGCM used in the IPCC Fourth Assessment Report. According to Giorgi (2006) southern Europe and the whole of the Mediterranean region are the so-called climate change "hot spots". This essentially implies that either the region's climate is very responsive to climate change or that the impact of climate change on the environment or human activity could be very pronounced. The complexity of the European climate is further emphasised by the magnitude of its interannual variability. Based on the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis dataset (ERA-40; Uppala et al. 2005), Monaghan and Bromwich (2008) concluded that interannual variability of the 2 m temperature (over land points only) is largest for Europe of all seven continents considered, and the second largest for mean sea-level pressure (MSLP). However, before focusing on climate change over the southern Europe, the results for global fields are discussed. Due to spatial and temporal interactions, the knowledge of climate change over the globe is essential, i.e. changes over a limited domain could be viewed only in association with changes on a much wider scale.

An analysis of the model systematic errors is also given. In the context of climate change—which usually invokes to making the difference between future and present climate—it may not seem obvious whether it is necessary to diagnose model systematic errors. However, as it will be discussed below, such an analysis is important when estimating the effects of climate change in various impact studies (e.g. extremes in short-term climate), when applying global model data in dynamical downscaling and in climate change detection and attribution studies.

In the following section, the model, data and methodology are explained. In Section 3, a brief discussion of model systematic biases, both global and regional, is presented. Section 4 deals with climate change on global scale, whereas in Section 5 the discussion of climate change is focused to the results for a limited area domain. In Section 6 summary and conclusions are given.

2 Data and methods of analysis

2.1 Global climate model

EH5OM is an AOGCM consisting of ECHAM5 atmospheric model and MPI-OM oceanic model. The ECHAM5 is the fifth generation of the ECHAM atmospheric general circulation model (GCM) with a spectral dynamical core defined for the purpose of climate integrations at the T63L31 resolution (Roeckner et al. 2003). A large body of literature dealing with various aspects of climate research based specifically on climate integrations with either EH5OM model, or with its atmospheric component ECHAM5, is available (also for earlier versions of the same model) – to name but few, e.g. Roeckner et al. (2006), Roesch and Roeckner (2006), Hagemann et al. (2006), Wild and Roeckner (2006), Müller and Roeckner (2006). The MPI-OM is the global ocean general circulation model with horizontal resolution of 1.5° and 40 vertical levels (Marsland et al. 2003). The model includes the calculation of sea ice and parameterisation of additional wind mixing. Frequency of coupling between the atmospheric and oceanic models is on the daily basis with no flux adjustment (Jungclaus et al. 2006).

2.2 Climate integrations and data used

Three integrations for the twentieth century climate (20C3M) and three integrations for the future climate (twenty-first century) under the IPCC A2 scenario (SRESA2) of the EH5OM model were at our disposal. For the twentieth century climate, the three 20C3M integrations differ among each other in the definition of initial conditions, i.e. they are separated by 25 years and were all taken from the so-called pre-industrial control experiment. The 20C3M integrations cover the period 1860–2000. The three SRESA2 experiments were simply the continuation of the three 20C3M runs and cover the period 2001 to 2100. The data for this study were retrieved from the Program for Climate Model Diagnosis and Intercomparison (PCMDI) archive of the Lawrence Livermore National Laboratory (LLNL) in Livermore, California, USA. They are part of a wider CMIP3 archive (see Meehl et al. 2007a), defined and set up by the World Climate Research Programme (WCRP).

To represent the twentieth century climate, the data from the period 1961–1990 are used, whereas the future climate has been estimated from the years 2041–2070. For the purpose of this study, the 1961–1990 modelling climate will be referred to as the "present climate". In terms of observational data some differences between the last decade of the twentieth century and the 1961–1990 average could be noticed. For example, the annual globally averaged temperature anomalies for many years from the last decade of the twentieth century (1991–2000) fall within the range of the warmest anomalies on record (e.g. Jones et al. 2001), and thus clearly deviate from the observed climatological means of the 1961–1990 period. The two 30-year periods (20C3M and SRESA2) represent a classical climatological time span required for representative, stable and statistically significant climatological samples of the two different populations (WMO 1967). It is sufficiently long to confidently extract climatological signals and compute variability of both climates.

The choice of period for future climate assessment is usually arbitrary. Meehl et al. (2007b) discussed three periods: an early-century (2011–2030), a mid-century (2046–2065) and the late century (2080–2099). In our study this choice is a compromising one: we did not want to go too far into the future, and believe that the selection of the years 2041–2070 would represent well the climate around the middle of the twenty-first century. For comparison purposes, the Meehl's et al. (2007b) mid-century period is closest to our choice, albeit 10 years shorter. For 20C3M simulations, the concentrations of carbon dioxide (CO₂), methane (CH₄) and nitrous oxide (N₂O) in EH5OM are specified at observed values (years 1860–2000),¹ whereas the anthropogenic ozone is defined as the difference between the actual and pre-industrial values. For future climate, the time evolution of the greenhouse gases (GHG) is based on the A2 emission scenario (Nakićenović et al. 2000; year-on-year concentrations are given at the same Web page as in footnote 1) with stratospheric ozone included. The model also includes the first indirect effect of aerosols. The simulation details of radiative forcing for future climate projections are given in Meehl et al. (2007b).

Most of the results are shown and discussed in terms of ensemble averages for the December to February (DJF) and June to August (JJA) seasons. In the mid-latitudes, many climate-related phenomena attain extreme values in these two seasons and it is important to assess whether and how they are modified by climate change. The results from individual runs are also discussed when additional clarification or relatively large variations within an ensemble are found.

For most parameters, the EH5OM model systematic errors are estimated by comparing with the reference climate as defined from ERA-40. The ERA-40 seasonal averages for the 1961–1990 period are defined at the same resolution as the EH5OM data, i.e. T63 (N48 Gaussian grid), although the original ERA-40 resolution is T159 (N80 Gaussian grid). For precipitation, the model results are compared against the Climatic Research Unit² (CRU) data (e.g. New et al. 2002). The CRU precipitation is defined over the land points only and was interpolated from the original $0.5 \times$ 0.5° regular latitude/longitude grid to the model N48 Gaussian grid. Relatively highresolution CRU verification data might be viewed as an inappropriate disadvantage for the model; however, the verification of precipitation discussed in the next section

¹For time evolution of GHG concentrations see http://www.cnrm.meteo.fr/ensembles/public/results/ results.html.

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is focused mainly on qualitative differences. Besides, the CRU precipitation covers the period 1961–1990, consistent with verification of other variables based on ERA-40 data. Other available global precipitation datasets cover shorter periods or periods that do not coincide with our present climate.

2.3 Method of analysis

For the assessment of climate change from the EH5OM model, the three simple and relatively crude measures are used. First we compute climatological means for both seasonal "climate" samples of the twentieth and twenty-first centuries. Since three model integrations were at our disposal, these climatological means include averaging over all model realisations. Such ensemble averages further enhance the relative importance of the climate signal because it eliminates, at least partly, some of uncertainties related to internal variability. The likely climate change is then revealed as the difference between ensemble averages.

Statistical significance of the differences between the means of the present and future climates is tested by performing the conventional Student t statistics, despite some criticism of the "null hypothesis significance testing" (e.g. Nicholls 2001). The issue of climate change seems to be too important to be left without any information on (statistical) reliability of the results. Although the size of our climatological samples might not be always adequate for statistical testing, the results of the t statistics complement those based solely on climate averages.

In addition to the above climatological means, interannual variability of both climates has been computed and compared. This is important, because irrespective of possible *mean* climate change detected from model integrations, the change of interannual variability may have a significant influence on behaviour of extreme climate events. Katz and Brown (1992) argued that climate models must detect changes in climate variability, and that assessments of future climate based solely on changes in means are not reliable. For both present and future climates, the interannual variability is derived by taking into account all individual model realisations.

3 Verification of the model present climate

In this section some aspects of EH5OM model systematic errors are briefly discussed. This important issue sometimes tends to be neglected in the climate change studies because most of them are normally focused on the differences between present and future climates. In such a context model errors are usually, by the analysis design, excluded (cancelled out) from the consideration. However, as discussed by e.g. Hegerl et al. (2006), detection and attribution analyses of the climate change signal are sensitive to uncertainties associated with model errors, in particular if climate models share common errors. The confidence in a climate model results may be, broadly speaking, determined by a ratio between the magnitude of model climate response and its systematic errors (e.g. Déqué et al. 2005). The smaller the errors the more confidence we might have that climate change projections by a given GCM are less influenced by modelling uncertainties.

For verification purposes, the EH5OM present climate (1961–1990) was derived from monthly means for the DJF and JJA seasons, averaged over three model

realisations. The observed climate, based on ERA-40 monthly mean and CRU data, is derived for exactly the same period. For brevity, the focus is on the time-mean errors; however, some other aspects of error statistics are also discussed.

3.1 Model errors on global scale

The largest upper-air temperature errors in EH5OM integrations are found in the polar upper troposphere and lower stratosphere. Zonal averages reveal a strong cooling in both hemispheres exceeding -12° C at 200 hPa in the summer hemispheres, and about a half of this magnitude in the winter hemispheres (Fig. 1a is for JJA only). In the middle and lower atmosphere, temperature errors are much reduced when compared to a strong stratospheric cooling. The error pattern, similar to that depicted in Fig. 1a but with smaller amplitude, has been discussed in some detail by Roeckner et al. (2006), who examined errors from various configurations of the ECHAM5 model. Similarity and persistence of errors between EH5OM (Fig. 1a) and ECHAM5 (Roeckner et al. 2006) are evident despite the different experimental set-up (EH5OM is a fully coupled atmosphere-ocean model whereas ECHAM5 was forced by observed sea surface temperatures), the different verification datasets (ERA-40 vs. ERA-15) and the different verification periods considered (1961–1990 vs. 1977–1993).

The above temperature error pattern in the upper atmosphere has been noted in many early climate models (e.g. Boer et al. 1992). Johnson (1997) attributed the "coldness" to the failure of climate models to simulate accurately the meridional



Fig. 1 Differences between EH5OM ensemble mean and ERA-40 for **a** JJA zonally-averaged temperature, **b** JJA zonally-averaged *u*-wind, **c** DJF u200, and **d** JJA T200 interannual standard deviation. Contours every 1° C in **a**, every 1 ms^{-1} in **b** and **c**, and every 0.2° C in **d**

energy transport due to spurious dispersion and diffusion. Clearly, many years since this type of error pattern was detected, it is still evident in climate simulations, although its magnitude has been largely reduced when compared to Boer et al. (1992). The reduction of error amplitude is due to a continuous improvement of climate models' dynamics and physics (e.g. the inclusion of stratospheric ozone).

Strong gradients in zonal mean temperature errors, as depicted in Fig. 1a, are associated with relatively large errors in zonal mean zonal wind. The *u*-wind errors are largest in the stratosphere and extend down into the troposphere, indicating their mainly barotropic nature (Fig. 1b is for JJA only). They can be interpreted as a strengthening of the subtropical jet cores in the model relative to ERA-40 (westerly bias), together with a displacement of the maximum wind axis in the troposphere. The *u*-wind error pattern and amplitude in Fig. 1b is again very similar to that shown in Roeckner et al. (2006). The exception is an increased positive error in the equatorial upper troposphere and lower stratosphere of both seasons in Roeckner et al. (2006); this might be attributed, at least partly, to insufficient vertical resolution in ERA-15 (as discussed in Branković et al. 2002).

The spatial extent of the mid-latitude westerly bias is clearly seen in Fig. 1c. This is a fairly robust deficiency of the winter circulation in many GCMs. In the northern hemisphere, Europe is most exposed to a stronger than observed westerly winds (also known as zonalization, see for example, Tibaldi et al. 1997), which penetrate deep into the continent from the adjacent Atlantic Ocean. A typical implication of such a model deficiency is a larger than observed precipitation, in particular over parts of western Europe (not shown).

In addition to errors in climatological means, Fig. 1d shows an example of model errors in reproducing the 1961–1990 observed interannual variability for the summer temperature at 200 hPa, T200. Errors are expressed in terms of the difference in standard deviation between the present climate and ERA-40. In the mid- and high-latitudes, interannual variability in the model is underestimated with respect to analysed values, whereas in the tropics it is overestimated. Similar error, but with reduced amplitude, extends throughout the atmosphere. The errors shown in Fig. 1d are by no means typical of all parameters, levels and seasons, and they can vary considerably at different geographical locations. For wind, for example, errors in variability are larger at the jet level and in the low-latitudes than at lower levels or in mid- and high latitudes.

For temperature closer to surface (T2m), an erroneous warming in both winter hemisphere high latitudes is seen (Fig. 2a, b). The largest positive errors are found over the oceans during winter (more than $+10^{\circ}$ C over the Southern Ocean) when these oceanic regions are covered by sea ice. The erroneous warming in Fig. 2a, b indicates an insufficient cooling by the model sea ice, which is confirmed by a different positioning of the 0°C isoline in the model and in the ERA-40 data (see Kållberg et al. 2005). Lambert and Boer (2001) noted that such a temperature error pattern is typical for the coupled models with no flux adjustment at the air-sea boundary. Covey et al. (2003) found that the range of observed global annual mean surface air temperature (with all observational uncertainties included) is between 13.5°C and 14.0°C, but the range in 16 coupled models from CMIP2³ was between 11.5°C and 16.5°C.

³The JSC/CLIVAR Coupled Model Intercomparison Project (Meehl et al. 2000).



Fig. 2 Differences between EH5OM ensemble mean and ERA-40 for T2m in **a** DJF, **b** JJA; zonally averaged total precipitation over land points only for EH5OM ensemble mean (*blue*) and CRU interpolated to the model grid (*red*) for **c** DJF, and **d** JJA. Contours at 1, 2, 4, 6, 8, 10°C in **a** and **b**

Another prominent feature in Fig. 2a and b is the warming off the west coasts of Africa and both Americas. The pattern and the extent of this model error coincide with the climatological coverage of the oceanic low stratiform cloud decks (see, for example, Klein and Hartmann 1993). Lambert and Boer (2001) diagnosed similar error in many CMIP1 coupled models without flux correction. They noted that such errors are "likely the result of a lack of simulated low-level cloudiness". This seems to be the case with the EH5OM model as well, where the amount of the Peruvian and Namibian stratus clouds in DJF is reduced with respect to ERA-40 by 30% and 20%, respectively (not shown). However, for the Californian cloud deck this argument may not be valid, since model error in the cloud field there is less than 10%.

Figure 2c and d shows that the EH5OM model underestimates tropical precipitation. A relatively large difference around 20° N in JJA is related to a poor representation of the south Asia monsoon in the model, whereas the peak at about 27° N indicates an overestimation of precipitation over the Himalayas (not shown). In the northern extratropics, the model tends to overestimate precipitation regardless of the season. These results are consistent with those for 18 coupled models discussed by Covey et al. (2003).

3.2 Model errors on regional scales

The north Atlantic westerly bias in wind shown in Fig. 1 is associated with an erroneously reduced MSLP in a wide swath over parts of Europe. In summer, the negative MSLP error extends northward from 50° N (Fig. 3a); in winter, it moves a little southward (not shown). Thus, irrespective of season, the positive MSLP error covers southern Europe and the Mediterranean region. Such an error would



Fig. 3 a Difference between EH5OM ensemble mean and ERA-40 for JJA MSLP; **b** JJA total precipitation over land points only averaged over $(35^\circ-50^\circ \text{ N}, 10^\circ \text{ W}-30^\circ \text{ E})$ for CRU data at the original $0.5 \times 0.5^\circ$ grid (*solid squares*), ensemble averages (*red circles*) and individual model realisations (*crosses*). Contours in **a** every 0.5 hPa

normally implicate a reduced model precipitation over these areas. Indeed, Fig. 3b shows a persistent model underestimation of summer precipitation over southern Europe $(35-50^{\circ} \text{ N}, 10^{\circ} \text{ W}-30^{\circ} \text{ E})$ for the whole 1961–1990 period. Only in few, out of the total 90, individual model realisations (shown as crosses in Fig. 3b) does precipitation come close to the CRU observed values. Such a model underestimation of precipitation ought to be taken into account when addressing the effects of climate change over these regions.

A relatively strong westerly bias seen in Fig. 1c would have normally caused an increased precipitation in many parts of Europe. However, such an increase is mainly

confined to northern Europe north of 55° N (Scandinavia), whereas the rest of the continent exhibits a reduction in summer precipitation when compared with CRU data (not shown). Since in summer the dominant type of precipitation is generated through convection, it might be assumed therefore, that a reduction in summer precipitation is associated with a reduced or inadequate convective activity in the model.

4 Climate change in global fields

4.1 Temperature

The climate change in the temperature field is manifested as global warming through the entire troposphere (Fig. 4a, b). Irrespective of season, the largest temperature increase in future climate, amounting to about 4°C in zonal average, will occur in the tropical upper troposphere and the lower stratosphere. However, even stronger warming, in excess of 5°C, can be seen during winter at the lowest levels of the northern polar regions. Temperature differences shown in Fig. 4a and b are highly statistically significant. The pattern of differences displayed in Fig. 4a and b, particularly in the upper troposphere, is in a broad agreement with that from Covey et al. (2003) and with the mid-century period from Meehl et al. (2007b). It also shows a great deal of similarities with the results obtained by Boer et al. (2000)



Fig. 4 Ensemble-mean difference between future and present climate for **a** DJF zonally-averaged temperature, **b** JJA zonally averaged temperature, **c** JJA T850, and **d** JJA T850 interannual standard deviation. Contours every 0.5°C in **a**–**c**, and 0.1°C in **d**

for the CCCma⁴ model. A detailed comparison of Fig. 4a and b and the results from the above studies is not easy because different averages (seasonal vs. annual) and different periods are analysed. With the exception of the stratosphere and the upper troposphere, a large part of tropospheric warming shown in Fig. 4a and b is very little affected by model systematic error (cf. Fig. 1a and similar for DJF). This result is important, since it increases our confidence in the global temperature change projection as defined by EH5OM model.

The horizontal distribution of tropical differences at high tropospheric levels is fairly zonally uniform. However, at lower levels this uniformity is disrupted and some regions with a notably larger temperature increase emerge. In particular, this is the case in JJA for eastern North America, south-west Europe and western Mediterranean, south-west and central Asia, (Fig. 4c shows temperature at 850 hPa, T850). For the above regions, this excessive warming is extended to the surface and in zonally-averaged differences (Fig. 4b), it is seen as the difference "dip" at around 40° N. Also, the largest warming in the above regions is associated with relatively small systematic errors, below $\pm 1^{\circ}$ C. However, in some other (mostly oceanic) regions with a relatively large warming, model error exceeds 2–3°C (not shown).

The change in interannual variability follows the pattern of change in climatological means: it decreases from the upper troposphere towards the surface. On average, there will be an increase in the variability of temperature in future climate relative to the period 1961–1990 (see Fig. 4d for T850). Generally, the change of variability is more pronounced in JJA than in DJF. Figure 4c and d indicates that in many regions increased variability coincides with an increase in the means (e.g. in parts of North America, south Europe and the Mediterranean, east Africa and the Arabian Peninsula, South America, southern Africa). However, in terms of amplitude, changes in variability are generally much smaller than the time-mean warming. For the above regions, the difference in standard deviation is smaller than the difference in climatological means for approximately an order of magnitude. Räisänen (2002) found similar results for T2m analysed from 19 CMIP2 models and suggested that future changes in the extremes of interannual temperature variability will largely depend on the magnitude of time-averages.

One possible reason for change in interannual variability shown in Fig. 4d could be a change in the mean state and interannual variability of El Niño-Southern Oscillation (ENSO) events in future climate, which in turn would induce a change in remote response of the model atmosphere. For the 30-year periods considered here, the DJF area-averaged SST in the Niño3.4 region increased by 1.6°, from 25.7°C in present climate to 27.3°C in future climate. This is consistent with the twenty-first century change of the Niño3.4 index, as demonstrated by Müller and Roeckner (2006) for the IPCC A1B emission scenario and for a century-long period. However, since we did not have the EH5OM control integrations at our disposal, the computation of the SST interannual variability in the equatorial Pacific yields somewhat different results when compared to those from Müller and Roeckner (2006): expressed in terms of standard deviation derived from 90 values (30 years times three model integrations), the SST interannual variation has remained unchanged in future climate for the Niño3.4 region (1.76).

⁴Canadian Centre for Climate Modelling and Analysis, University of Victoria, BC, Canada

For the near-surface temperature (T2m), the consistency of zonally averaged climate differences among individual EH5OM runs are shown in Fig. 5. Although some variation in the three runs is seen, a general pattern of the differences common to all model integrations is apparent. In other words, the uncertainty of our results as revealed by the variation of zonal averages of multiple model realisations is much smaller than the magnitude of climate change. Indeed, the differences among multiple realisations as depicted in Fig. 5 could be considered relatively small in view of the fact that Barnett (1999) has shown that the differences between future and present climate among *various* climate models could be much larger than those between the three runs discussed here.



Fig. 5 Zonally-averaged differences between future and present climate for the three individual EH5OM runs for T2m in a DJF, and b JJA

A striking feature in Fig. 5 is that, for all latitudes and in all runs, the IPCC A2 emission scenario for years 2041–2070 produces no negative or even zero differences. The warming is largest in the winter hemispheres, poleward of 60°, in particular in the northern winter (note the different *y*-axes in Fig. 5a, b). Shindell et al. (1999) attributed such a warming to the strengthening of zonal winds around the Arctic, associated with an increase in the concentration of GHG. They claim that, although the warming trend occurs through a natural variation (Arctic Oscillation), it could be induced by anthropogenic impact and may continue to rise. The DJF warming north of 60° N shown in Fig. 5a is larger than zonally averaged T2m errors with respect to ERA-40 or CRU verification datasets (not shown). This result again confirms that the climate change signal for the near-surface temperature is very strong indeed; this is consistent with the findings of some other authors as well (cf. Déqué et al. 2005). In summer, zonally averaged warming is much smaller than its winter counterpart and it is comparable to the model systematic error.

4.2 Global circulation

The strongest change in zonally averaged *u*-wind in stratospheric mid-latitudes (Fig. 6a, b) is associated with the largest meridional gradient in temperature (cf. Fig. 4a, b). The increase of *u*-wind is related to an intensification of winter jet cores at both hemispheres from a little above 40 ms^{-1} in present climate to more than 45 ms^{-1} in future climate (an increase of about 10%; not shown). In the southern



Fig. 6 Ensemble-mean difference between future and present climate for **a** DJF zonally-averaged *u*-wind, **b** JJA zonally-averaged *u*-wind, **c** DJF u200, and **d** DJF u200 interannual standard deviation. Contours every 0.5 ms^{-1} in **a**, **b** and **d**, and 1 ms^{-1} in **c**

hemisphere, this jet core intensification is also associated with a slight poleward shift of wind maxima, thus effecting a strengthening of *u*-wind throughout the troposphere. Almost all changes in Fig. 6a and b are statistically significant at the 95% confidence level, indicating a stable and consistent signal in climate integrations. Similar result is obtained in most models that took part in the latest IPCC assessment (Meehl et al. 2007b).

The strengthening of jets in both hemispheres in future climate - on average more than 4 ms⁻¹ over the north Pacific and the Southern Ocean—is clearly seen for DJF in Fig. 6c. The prevailing westerlies over western Europe are also strengthened in future climate and, according to Pinto et al. (2007), will result in an increased cyclonic activity. Based on the results from EH5OM model, Müller and Roeckner (2006) link the future change in the northern hemisphere mid-latitude flow, in particular over the Atlantic, to an increase of the tropical Pacific SSTs. In the southern hemisphere winter (JJA), the strengthening reaches even more than 6 ms⁻¹ south of Tasmania (not shown; see also Stone and Fyfe 2005). However, the interpretation of this strengthening should be considered carefully because of the erroneous wind zonalization in EH5OM model, as discussed in Section 3. For example, the strengthening of the winter southern circumpolar vortex (Fig. 6b) partly coincides with the model underestimation of zonal winds at 60° S (Fig. 1b).

The difference in standard deviation of the DJF 200 hPa wind magnitude (Fig. 6d) is a good overall measure of the change in interannual variability for both tropics and extratropics. In many places the difference maxima in standard deviation coincide with the difference maxima in the mean field. In future climate, variability of wind at the jet level will experience a large increase in the eastern Pacific and subtropical Atlantic, but less so in the north Atlantic and European regions. The largest decrease in variability is seen in the tropical Atlantic, the bordering region of the subtropical African jet.

4.3 Precipitation

The global mean precipitation will be increased in future climate, but, as revealed by Fig. 7, there are also latitudinal belts where precipitation is reduced. Zonal averages in Fig. 7 are computed for land points only in order to be consistent with the discussion on precipitation verification (Fig. 2c, d). The largest increase of about 0.4 mm day^{-1} will occur in summer seasons in tropical bands associated with the rising branch of the local Hadley circulation: in the northern hemisphere, the increase is found over the equatorial Africa and south Asian monsoon regions, whereas in the southern hemisphere it is related mainly to Amazonia (see also Fig. 8a, b). However, the above results must be viewed in the context of systematic underestimation of tropical rainfall by the model as shown in Fig. 2c and d.

Although in Fig. 7 the overall differences among individual model realisations may look comparatively small, and a great deal of consistency for the three runs does exist, some regions could be identified where the spread among individual runs exceeds the induced climate change. These include, for example, latitudinal belts around 15° S in DJF or around the equator and close to 30° N in JJA. This is in contrast with the T2m zonal averages in Fig. 5 where the differences among individual model runs were smaller than the climate change. In addition, a comparison of zonal averages in Fig. 7 and model systematic biases in Fig. 2c and d indicates that the climate change signal



Fig. 7 Zonally-averaged differences between future and present climate for the three individual EH5OM runs for total precipitation over land points only in a DJF, and b JJA

for total precipitation is much weaker than that for the near-surface temperature (Fig. 5).

The spatial variation of precipitation changes (Fig. 8a, b) in the tropics is partly due to an increase in the future precipitation (between 5% and 10%) and partly to a redistribution of precipitation within the inter-tropical convergence zones (ITCZs). A tendency of the equatorial Pacific SST anomalies to be warmer in future climate (as discussed in Section 4.1, see also Müller and Roeckner 2006), may cause, via deep convection, non-negligible nonlinear adjustment into the equatorial climate system. Chou et al. (2006) also found, for an earlier version of the same model



Fig. 8 Ensemble-mean difference between future and present climate for total precipitation in **a** DJF, **b** JJA, **c** JJA *t* statistics at the 95% confidence level, and **d** JJA coefficient of variation of precipitation. Contours every 0.2, 0.5, 1, 2, 5, 10 mm day⁻¹ in **a** and **b**, and 0.1, 0.2, 1, 2, 5, 10 in **d**

(ECHAM4/OPYC3), that the largest precipitation climate changes are occurring in the regions of a relatively strong convection. In JJA, climate changes are statistically significant at the 95% confidence level (Fig. 8c), less so in DJF. This is in contrast with the Déqué and Royer (1991) results, who found stronger climate sensitivity in their (perpetual) January than in July experiment; however, such a discrepancy may arise because of different model parameterisations and experimental set up. In some regions, the differences in precipitation averages change sign between the opposing seasons (e.g. Europe, eastern Asia, central Brazil). However, in many others (southeast US, eastern Canada, the Gulf of California, northern Siberia, southeast Australia) the precipitation change retains the same sign in both JJA and DJF. Of course, such behaviour might induce, depending on the region and the sign of changes, a beneficial or detrimental effect on human life and activities.

The changes in interannual variability of precipitation, in terms of the coefficient of variation (see for example, von Storch and Zwieres 1999), are shown for JJA only (Fig. 8d). The largest changes, both positive and negative, could be generally linked to the regions with relatively small precipitation amounts. The largest positive changes, indicating an increase in the variability of precipitation, are found, for example, in the subtropics, the Mediterranean and in the eastern tropical Pacific. For the latter, an increase in variability is consistent with a shift of the EH5OM tropical Pacific SST anomalies towards higher positive values in the twenty-first century described by Müller and Roeckner (2006).

5 Climate change over southern Europe

Although the knowledge and understanding of global climate change is a prerequisite for a proper assessment of climate changes on regional scales, the pattern, amplitude, variability and other attributes of global climate change cannot be simply extended to every corner of the globe. They will be influenced and modulated by features affecting local climate of the place or region considered. It has been shown in the previous section that climate change is by no means spatially uniform, i.e. various parts of the world will be affected in different ways (see also Giorgi and Bi 2005a, b). Some aspects of large-scale circulation (as revealed by temperature, wind, precipitation) over a wider European region will experience a moderate to a large change in the twenty-first century. In this section, we focus to the effects of climate change derived from the EH5OM ensemble surface or near-surface fields over southern Europe.

Because of EH5OM's relatively coarse horizontal resolution, interannual variability over a limited domain may not be representative for all parameters, in particular for those with discreet and incoherent characteristics (like, for example, precipitation). For example, Good and Lowe (2006) have demonstrated that local trends in precipitation can be much larger than or even opposite to the large-scale regional averages. This supports the argument that dynamical downscaling with a limited area model might be a better way to study climate change on regional scales. In a companion paper (Branković et al., in preparation), the results of dynamical downscaling for the southern Europe with a regional climate model forced by the EH5OM global data are discussed and compared with those from the global model presented in the current study.

5.1 Near surface temperature (T2m)

It has been demonstrated in Section 4.1 that under the IPCC A2 radiative forcing a globe-wide warming throughout the depth of the atmosphere will occur in future climate. In particular, southern Europe and the Mediterranean were identified as the regions where the temperature increase will be larger than in many other parts of the globe. For near-surface temperature, T2m, such a warming is also seen (Fig. 9a, b; see also Christensen et al. 2007). Here, the summer season stands out with an excessive warming of more than 4°C over the Iberian Peninsula and more than 3.5°C over the northern Adriatic and the adjacent land regions. A comparison of the T2m climate change derived from individual integrations yields a consistent and coherent model response, i.e. a very little difference in the spatial temperature variations are seen from one model run to the other (not shown, but cf. Fig. 5b for zonal averages). Thus, similar to the result in Section 4.1, uncertainties due to inherent atmospheric variability are overwhelmed by the strength of the climate change signal. The intensity of the above signal is further confirmed by the t statistics, which is significant even at the 99% confidence level. Such a strong warming in seasonal mean would inevitably contribute to an increase of temperature extremes and more frequent heat waves or hot spells in summer over southern Europe and the Mediterranean (cf. for example, the results from Barnett et al. 2006; Clark et al. 2006; Tebaldi et al. 2006). Although in JJA a cold bias of slightly below -1° C prevails



Fig. 9 Ensemble-mean differences between future and present climate for T2m in **a** DJF, **b** JJA; for T2m interannual standard deviation in **c** DJF and **d** JJA. Contours every 0.5 °C in **a** and **b**, and every 0.1°C in **c** and **d**

over parts of central and southern Europe (Fig. 2b), and consequently the above result might be viewed with some caution, the large amplitude of warming in Fig. 9b overpowers this model deficiency.

In DJF, the warming is more pronounced over the north-eastern Europe, where in present climate normally colder, below freezing, temperatures prevail. The temperature increase between 2.5°C and 3°C indicates that climate change in winter may have a more dramatic overall impact on north-eastern Europe (in particular for snow cover, see Section 5.3 below), than on the Mediterranean or western Europe regions. According to Rowell (2005), a robust east-west gradient across central Europe in the winter warming (Fig. 9a) is a combination of relatively modest warming of the Atlantic in the west and a weakened snow-albedo feedback in eastern Europe, which is mainly responsible for low temperatures in present climate. The near-surface temperature increase in winter is associated with very little or almost no change in average cloudiness (not shown). A possible explanation for such a small change in cloudiness could be that during the winter season climatological amount of clouds in the north-eastern Europe is very high in any case—for example, more than 80% in the regions where the temperature difference in Fig. 9a exceeds 2.5°C. Thus, over the north-eastern Europe the impact of radiative forcing seems to have a little effect on the relationship between clouds and near-surface temperature during winter. Over the Mediterranean, where the winter climatological cloud coverage is below 60%, the reduction in cloudiness in future climate is more obvious (not shown).

The change of interannual variability in DJF over southern Europe and the Mediterranean is nearly negligible (Fig. 9c), in spite of a substantial increase in the mean T2m in future climate. In contrast, in JJA (Fig. 9d), a widespread increase in interannual variability of T2m is seen. In addition, there is a tendency that maxima in the difference of standard deviation coincide with maxima in the difference of the means. However, similar to what was discussed in Section 4.1 for upper-air fields, the change in amplitude of interannual variability is (much) smaller than the actual climate change. Therefore, the argument put forward by Räisänen (2002) holds for the EH5OM near surface temperature over the limited domain considered. An increase in interannual variance of the summer T2m could be associated with a strong drying over land (Rowell 2005; cf. Fig. 11b). In increased dry conditions, a coupling between land surface (with reduced soil moisture, Fig. 12d) and the atmosphere above will become stronger and hence positive feedback will enhance the variation of year-to-year anomalies (Seneviratne et al. 2006).

5.2 Mean sea level pressure (MSLP)

Whilst the climate change over the Mediterranean and southern Europe implies an increase in MSLP during DJF, the opposite sign is seen in JJA (Fig. 10a, b). The JJA t statistics indicates that the difference between future and present climate is significant at the 95% confidence level (Fig. 10c; similar result holds for DJF, but with the 95% confidence level shading shifted southward). However, as depicted in Fig. 3a, an error with the amplitude of about two to three times larger than the associated climate change may have an adverse effect on interpretation of the above results. Since in both summer and winter, the MSLP error pattern over the domain of interest is very similar (see Section 3.2), errors in MSLP might affect the respective climate change in Fig. 10a and b in different ways.

The increased pressure in the future southern Europe and Mediterranean winters indicates an increased frequency of or more prolonged anticyclonic weather types over the region; or likewise, a decrease in the frequency of or the shortening of cyclonic situations. This would eventually lead to more stable winters than they were in the period 1961–1990, which in turn might affect some other parameters, like for example, cloudiness and precipitation. Despite such a "stabilisation" in the climatological mean, the interannual variation of MSLP during winter will be increased in future climate over much of Europe (Fig. 10d). It may be possible that such an increase in variability represents a combination of large-scale and local effects. Positive difference over southern Europe in Fig. 10d may be viewed as an eastward extension of the increased north-eastern Atlantic variability in MSLP (not shown, but seen partly in Fig. 10d over the Bay of Biscay), associated with changes in the North Atlantic Oscillation (NAO). Namely, the mean change in winter MSLP (Fig. 10a) bears the signature of a tendency towards the positive phase of the NAO in future climate (enhanced westerlies over the North Atlantic and a northward shift of the storm track; see e.g. Ulbrich and Christoph 1999). On the other hand, a tendency of decreasing snow amount and snow extension in future climate (see the next



Fig. 10 Ensemble-mean differences between future and present climate for MSLP in a DJF, b JJA, d DJF interannual standard deviation and c JJA t statistics at the 95% confidence level. Contours every 0.2 hPa in **a** and **b**, and every 0.1 hPa in **d**

subsection) may introduce a larger variability in coupling between the atmosphere and changes in underlying (less) cold surface over the south-eastern Europe.

5.3 Precipitation

5.3.1 Total precipitation

Whereas in DJF a decrease in precipitation is seen mainly over the Mediterranean Sea and some surrounding coastal regions, in JJA it is centred over the continental Europe and affecting only the northern Mediterranean (Fig. 11a, b). In both seasons, the amplitude of the precipitation reduction exceeds a little more than 0.5 mm day⁻¹, or about 45 mm per season. For north Africa in DJF and western Europe in JJA, this represents a substantial decrease, between 25% and 30% of total precipitation, respectively. Both summer and winter reductions in precipitation are statistically significant at the 95% confidence level (c.f. Fig. 8c). Associated with the warming of more than 3°C (see Section 5.1) such a reduction in summer precipitation may seriously alter the future European climate. Our results for Europe and the Mediterranean are in good agreement with those of Voss et al. (2002) obtained for an earlier version of the same model (ECHAM4) at a higher horizontal resolution (T106) and with



Fig. 11 Ensemble-mean differences between future and present climate for total precipitation in **a** DJF, **b** JJA, **c** DJF total precipitation coefficient of variation, and **d** for DJF snow amount. Contours at 0.1, 0.2, 0.3, 0.5, 1, 2 mm day⁻¹ in **a** and **b**, at 0.05, 0.1, 0.2, 0.5 in **c**, and 0.5, 1, 2, 3, 5, 10 mm in **d**

Giorgi and Coppola (2007) who analysed 22 CMIP3 models under the A1B IPCC scenario. However, following the discussion in Section 3.2, an important *caveat* must be borne in mind: precipitation estimates in climate models are strongly nonlinear and corresponding errors may not always be removed by subtracting two climates. Hence, climate change of precipitation in the presence of non-negligible errors as indicated here may include a (significant) degree of uncertainty.

An increase in total precipitation in future climate is seen only in winter approximately north of 45° N. It is in good agreement with the results obtained by Rowell (2005), who attributes such an increase to a moister atmosphere in a warmer climate (see also Douville et al. 2002). The maximum increase, located in Western Europe, is associated with a strengthening of prevailing westerly winds extending from the Atlantic into the European continent (c.f. Fig. 6c). This increase in the winter precipitation is linked, however, with almost no change in precipitation interannual variability (Fig. 11c). The maximum increase of the winter variability is found, like in Rowell (2005), further south, over the Mediterranean Sea. Giorgi and Coppola (2007) obtained similar result for the precipitation coefficient of variation in 22 CMIP3 models. In summer, however, they found that whilst the maximum of variability remains located over the Mediterranean, an increased variability in precipitation extends to central Europe.

5.3.2 Snow

Figure 11d indicates a large decrease in surface snow over central and eastern Europe, in particular over the Alps and the Carpathian Mountains, and further away over Russia. This is consistent with a general reduction in the amount of snowfall (not shown). However, since in the above regions Fig. 11a indicates a small increase in total precipitation during DJF, a reduction in surface snow must be also a consequence of a general temperature increase. As discussed in Section 5.1, a reduction in surface snow also means a reduction in positive snow-albedo feedback, which will consequently bring about milder winter temperatures. In view of Fig. 11a, this will contribute to atmospheric conditions that favour the deposition of precipitation in the liquid form. In many places the reduction in surface snow is up to 50-60%, which is a little less than in Rowell (2005). The interannual variability of surface snow over the Alps, eastern Europe and Russia will also be drastically reduced in future climate by almost the same amount as the reduction in climatological average. Thus, based on changes in the mean and in interannual variability, surface snow for most of Europe will become a very uncertain parameter. This may affect the availability of soil moisture required in growing season.

5.4 Moisture

5.4.1 Low-level atmospheric moisture

Climate change of atmospheric moisture is analysed for specific humidity, q, at lowermost pressure levels. Irrespective of season, q will be larger in future than in present climate, with values increasing from the upper to the levels closer to surface, consistent with the notion that a warmer climate will induce a moister atmosphere (e.g. Manabe and Wetherald 1975). Similar to atmospheric warming, a moistening of the lower troposphere in future climate is a globe-wide phenomenon (not shown), but it exhibits some spatially non-uniform variations on regional scales. For example, at 1,000 hPa in summer, a moistening in q1000 of more than 1 g kg⁻¹ over the Mediterranean Sea, western and central Europe is seen, whereas relative minima (between 0.4 and 0.8 g kg⁻¹) over the Balkan Peninsula, Turkey, north Africa and Spain are found (Fig. 12b). These relative minima coincide with local maxima in the warming of the near-surface temperature (cf. Fig. 9b), i.e. the warmest T2m change effects the air at the nearest atmospheric levels to be less moist than in surrounding regions. In winter, the moisture increase in future climate is largest over the Mediterranean Sea and over the north-eastern part of the domain (Fig. 12a). For the latter, this largest moistening is associated with the strongest warming in T2m (cf. Fig. 9a), i.e. the winter situation is contrasting to that in summer. The changes shown in Fig. 12a and b are statistically significant at the 99% confidence level.

The moistening of the atmosphere by evaporation plays an important role over the Mediterranean Sea—it is relatively strong in summer because insolation reaches its peak in the annual cycle, and also in winter because the sea surface is warmer than the adjacent land. Apart from the Mediterranean Sea, variations in the moistening of the lower troposphere can be attributed to different processes, which depend on season considered. For example, relative summer minima in q1000, seen mostly over land areas surrounding the Mediterranean Sea and coinciding with peaks in T2m, signify that the actual evaporation (or more correctly evapotranspiration) is much smaller



Fig. 12 Ensemble-mean differences for specific humidity at 1,000 hPa in **a** DJF, **b** JJA, and for soil moisture in **c** DJF and in **d** JJA. Contours every 0.2 g kg⁻¹ in **a** and **b**, and every 20 kg m⁻² in **c** and **d**

than *potential* evapotranspiration. This implies that sources of moisture, including soil moisture (see the discussion below), are exhausted, i.e. no additional moisture is available for the atmosphere. Therefore, the largest increase in T2m does not contribute to further moistening of the atmosphere, but rather to its relative drying. In DJF over the north-eastern part of the domain, the increase in future T2m is largest (Fig. 9a) and associated with a slight increase in precipitation (Fig. 11a) - their combination is thus causing a higher q at 1,000 hPa and at the levels immediately above. In addition, the reduced snow cover over the north-eastern part of the domain (Fig. 11d) is causing that less moisture is "locked" in snow and therefore made available for the atmosphere.

5.4.2 Soil moisture

If seasonally averaged hydrological balance over land were to be maintained, then the increase in the low-level atmospheric moisture and in the near-surface temperature, as shown and discussed above, must induce a redistribution of moisture within the land-atmosphere system. This will be done primarily at the expense of soil moisture: Fig. 12c and d shows that in both seasons soil moisture will be decreased in future climate, thus, being consistent with the changes of other parameters that influence the atmospheric water budget (here, due to a relatively coarse model resolution, the contouring covers parts of the Adriatic). The reduction in soil moisture is largest in summer over Spain, about 16% of the field capacity, and up to 10% over the Alps; in winter, the overall reduction is comparably smaller because of a larger amount of soil moisture available. The decrease in soil moisture is statistically significant in both seasons. Rowell and Jones (2006) found that the drying of soil in central and southern Europe during summer accounts, through a positive feedback, for about 20% reduction of summer, mainly convective, precipitation. This process should also be viewed as an effect of reduced soil moisture during spring, which in turn is a consequence of reduced snow and enhanced evaporation in the warmer future climate. By comparing 15 climate models, Wang (2005) also found that the Mediterranean region will become drier in both seasons in future climate, with a high degree of consistency among the models considered. She concluded that the Mediterranean is a "hot spot" for agricultural drought in the CO₂-induced global warming.

6 Summary and conclusions

An analysis of climate change between the "present" climate (1961–1990) and the climate of the middle twenty-first century (2041–2070) under the IPCC A2 scenario, based on the three-member ensemble of the EH5OM climate model, is performed. The main focus is on some near-surface and surface parameters over south Europe, the region that has been categorised as potentially very sensitive to climate change—the so-called climate change "hot spot" according to the definition by Giorgi (2006). In order to better evaluate and understand climate change on the regional scale, a limited analysis of global climate change for the same periods was also carried out. In addition, both, regional and global climate changes are related to the EH5OM model systematic errors. The assessment of modelling errors is not often present in the papers on climate change, because climate change is usually discussed in terms of the difference between future and present climates, and in that case it is assumed that model errors are removed from the consideration. However, various uncertainties related to climate change on both global and regional scales could be linked to climate models' systematic errors (e.g. Hegerl et al. 2006).

Our analysis is focused on winter (DJF) and summer (JJA) seasons. We mainly consider ensemble averages; however, the model spread or intra-ensemble variations are quantified for some parameters by comparing individual model realisations. The statistical significance of the differences between the chosen samples of future and present climates is tested as well. The change of interannual variability in the two periods is also examined. This is important because the changes of some climate extremes could be more strongly affected by interannual variability than by the changes in the mean.

Some of the large-scale model errors can be summarised as the following. The largest temperature errors (of more than -12° C in zonal averages) are found in the summer polar stratospheres. This kind of error "lingers" around for many years now; however, it has no major effect on the low-level temperatures. Due to improvements of GCMs in the stratosphere (e.g. inclusion of stratospheric ozone), the error has been reduced in the recent generation of GCMs when compared to e.g. Boer et al.

(1992), but it has not been removed. The near-surface temperature is too warm over sea ice and in the regions with the oceanic stratiform cloud decks. The former is an error typical of AOGCMs with no flux adjustment; the latter is often attributed to the lack of low-level clouds in climate models. Tropical precipitation is underestimated in EH5OM, in particular in JJA, whereas in both seasons precipitation in the northern mid-latitudes is overestimated.

Over Europe, the largest model error is manifested as the wind zonalization, i.e. an erroneous increase in prevailing westerlies that blow from the northern Atlantic into the European continent. The error is found in both seasons, and an important consequence is a reduction of MSLP over Europe to the north of 50°N and an increase of MSLP over southern Europe and the Mediterranean. In JJA, this erroneously increased MSLP is associated with a reduction of the model convective precipitation, the prevailing form of precipitation in summer.

The differences between future and present climates reveal a globe-wide increase in temperature. The warming is statistically significant at the highest confidence level. It is relatively uniform in the upper troposphere where the amplitude in zonal average nears 4°C in the tropics. The temperature differences are decreasing toward the poles, thus inducing an increase of the wind maxima—the speed in jet cores is increased by 10% in both hemispheres and irrespective of season. Further down, more spatial irregularities are seen and some regions with a relatively larger warming emerge. The interannual variability in temperature will be generally also increasing in future climate, but such an increase is about an order of magnitude smaller than the warming in the time-mean. Thus, the main part of the temperature climate change will largely depend on the magnitude of time-averages (cf. for example Räisänen 2002).

The internal variability, as revealed from the differences among the multiple model realisations, for temperature is much smaller than the magnitude of climate change, in particular for the northern extratropics in winter. In addition, the warming is larger than the model systematic error, thus confirming the significance of the climate change signal. Such a result strengthens our confidence in the model's ability to credibly simulate climate change on the global scale. However, for precipitation the model response contains more elements of uncertainty.

Climate change for southern Europe is considered for surface and near-surface parameters. The most important points could be summarized as the following. In winter, an increase in MSLP will bring a "stabilisation" of the region's future climate or, in other words, the prevailing "anticyclonic" circulation will affect the winter weather. A very strong model mean response is found for T2m in both seasons. The north-eastern Europe will be more affected in winter (with the maximum amplitude of over 3° C), and south Europe in summer (over 3.5° C). Since an increase in interannual variability that will occur in summer is much weaker than the change in the mean, it could be concluded that, similar to the upper-air temperature, an overall increase in the near-surface temperature extremes would be mainly due to the warming of the mean future climate.

Precipitation will decrease in summer over central and southern Europe. Regardless of a general increase in the winter precipitation over the continental Europe north of approximately 45°N, the increased near-surface temperature will cause a reduction in the amount of snow in north-eastern Europe. Irrespective of season, the low-level atmospheric moisture will increase, but soil moisture will decrease. The reduction in soil moisture over the southern Europe during summer would enhance unfavourable drought-like conditions for the agriculture.

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