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**Partners and research lines contributing to the deliverable**

The importance of assessing extremes is often emphasized because major events can cause loss of life, damage to infrastructure and large socio-economic impacts. It is crucial working towards the understanding of the physical mechanisms related with the occurrence of extremes and assessing the significance of any apparent changes in extremes over the long periods of time. Diverse partners in ENSEMBLES have been working on different aspects related with extreme events. In particular, of interest for the present deliverable four research lines have been developed as introduced below.

**Research line 1.** INGV have presented a methodology to study the interannual variability associated with summertime months in which extremely hot temperatures are frequent. Daily timeseries of maximum and minimum temperature fields (Tmax and Tmin respectively) are used to define indexes of extreme months based on the number of days crossing thresholds. An empirical orthogonal function (EOF) analysis is applied to the monthly indexes. EOF loadings give information about the geographical areas where the number of days per month with extreme temperatures has the largest variability. Correlations between the EOF principal components and the time series of other fields allow plotting maps highlighting the anomalies in the large scale circulation and in the SSTs that are associated with the occurrence of extreme events. The methodology is used to construct the “climatology” of the extremely hot summertime months over Europe. In terms of both interannual and intraseasonal variability, there are three regions in which the frequency of the extremely hot days per month homogeneously varies: north-west Europe, Euro-Mediterranean and Eurasia region. Although extremes over those regions occur during the whole summer (June to August), the anomalous climatic conditions associated with frequent heatwaves present some intraseasonal variability.

Extreme climate events over the north-west Europe and Eurasia are typically related to the occurrence of blocking situations. The intraseasonal variability of those patterns is related to the amplitude of the blocking, the relative location of the action centre and the wavetrain of anomalies downstream or upstream of the blocking. During June and July, blocking situations which give extremely hot climate conditions over north-west Europe are also associated with cold conditions over the eastern Mediterranean sector. The Euro-Mediterranean region is a transition area in which extratropical and tropical systems compete, influencing the occurrence of climate events: blockings tend to be related to extremely hot months during June while baroclinic anomalies dominate the variability of the climate events in July and August. Climate model simulations are able to capture the extreme-related variability in July, and climate change projections indicate that the more sensitive location for changes in extremes is north-west Europe.

**Research line 2.** AUTH group has been working on extreme temperature events over the Eastern Mediterranean region. Changes in extreme temperature over the Greek area are examined in relation to the geopotential thickness field (1000-500) hPa, for the period 1958-2000 (circulation type classification). A general increase (decrease) in frequency was found for the anticyclonic (cyclonic) circulation types. There is an increase (decrease) in the Tmax (Tmin) over the Greek area, with spatial and seasonal...
variations. A strong warming of the northern hemisphere in the recent decades is well established, which for Greece is evident in the summer mean but not in winter. In the results presented (Maheras et al., 2006, Good et al., 2007), the study period was dominated by warming in June, July, August and also September. This can be explained by decreasing trends in the frequency of the circulation types mostly responsible for the Etesian winter and an increase in the Mb type (circulation type with slack pressure field), both leading to more calm days. In addition there was a positive trend of the frequency of all anticyclonic types during summer leading to increased westerly/southwesterly flow, advecting warm air masses from Northern Africa. The observed trend of decreasing winter T\text{min}, accompanied by a trend of decreasing winter T\text{max}, contributes to a clear cooling trend. This overall decreasing tendency may be partially attributed to the general increase in frequency of surface anticyclonic types, which is related to northerly, northwesterly or northeasterly airflows over the study area.

Concerning the large scale atmospheric circulation, as a significant component of the natural climatic variability, the teleconnections play an important role in changes of the regional climate, and especially of extreme events. The impact of an upper level teleconnection pattern, called Eastern Mediterranean Pattern (EMP) on the extreme temperature regime of Eastern Mediterranean is investigated during winter. For the present climate, daily station data of maximum and minimum surface temperature data are employed for the period 1958-2000. For the future climate, datasets of the same parameters derived from the Regional Climatic Model HadRM3 for the period 2070-2100, using the IPCC emission scenario B2 for the evolution of the future atmospheric concentrations of greenhouse gases. The investigation of the impact is based on the Regularized Canonical Correlation Analysis (RCCA) while qualitative estimations of the composite temperature anomalies are performed for each phase of EMP. It was found that the EMP affects indeed the extreme events with inverse impacts between the two phases. The highest anomalies are found over continental Greece. More specifically, a positive phase of EMP is associated with a temperature decrease while the opposite occurs for the negative phase. In the future, the present impact according to each phase persists and intensifies, due to the estimated future shift of the EMP poles.

Research line 3. NERSC has leaded the research line on the Arctic melt seasons. The Arctic melt season has changed throughout the last century together with global warming. Melt season length has huge sociological, economical and ecological impact for Arctic. Extreme long melt seasons are examined with extreme value theory with global mean temperature as a covariate. The return period of events recognized as very extreme today will change with global warming. Today, the chance of having a melt season with duration 180 days or more is estimated to be 0.1%. With a global warming of 2° Celsius, the chance has increased to 24%.

Research line 4. People of KNMI have been working on identifying large-scale circulation structures that favour warm days in the Netherlands. They introduced a new statistical method that optimally links local temperature extremes to large-scale atmospheric circulation structures, and they applied it to an ensemble of climate model simulations over the period 1950-2100 under the SRES A1b scenario with the ECHAM5/MPI-OM coupled GCM. Daily July-August streamfunction fields at 500 hPa over the Euroatlantic region are used to identify large-scale circulation structures
that favour warm days in the Netherlands. Two patterns are identified (called Extreme Associated Functions); the most important one corresponds to a blocking high pressure system leading to subsidence and calm, dry and sunny conditions. The second EAF corresponds to a rare, easterly flow regime bringing warm, dry air into the region. The patterns simulated by the model compare very well to the patterns identified in the ECMWF reanalysis dataset (ERA40) over the 1958-2000 period. Over the future period 2071-2100 the same patterns are still responsible for the warm extremes and their probability of occurrence has not changed significantly. However, the corresponding local temperature extremes have warmed due to an increased net solar flux at the surface (due to reduced cloud cover) and a decreased evaporative cooling due to reduced soil moisture and more so for the stronger extremes. The same analysis is applied to eight more sites in Europe (France, Spain, Greece, Romania, Russia, Poland, Scandinavia and England). Only for the southernmost locations (Spain, Greece, Romania) do changes in the probability of occurrence of the EAFs contribute to the simulated changes in the local temperature extremes.
**Description of work and results archived**

**Research Line 1,**

*Heatwaves in Europe: Links with the regional to large-scale anomalies and climate change signal as reproduced by the SXG model*

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**RL1.1. Introduction**

As Europe is warming up (e.g., Jones and Moberg 2003, Giorgi 2002), there is evidences that extreme events might become more frequent and intense (Klein Tank and Können 2003, IPCC 2001, Easterling et. al 2000). The summer decade 1994-2003 has been the warmest period in Europe during the 20th century and the summer of 2003 has been the warmest summer since 1500 (Luterbacher et al., 2004). Moreover, some studies have documented an increase in the average frequency of heat waves of 0.24 per decade since 1880 (Della-Marta et al. 2007) while others, have reported the increase (decrease) of warm (cold) extremes during the last 40 years (Klein Tank and Können 2003, Yan et al. 2002). Some authors claim that extremes such as those recorded during the summer of 2003 do not fit into the statistics spanned by the observations of the last decades. Nevertheless, model simulations indicate that extraordinary hot summers could become more frequent in a warmer climate, when relevant changes in the variance of the temperature distribution are expected (Beniston 2004, Schär et al. 2004, Stott et al. 2004.).

Some studies relate the heatwaves to blockings in the extratropical circulation, altering the pattern of horizontal advections, warming the atmospheric surface levels due to adiabatic heating by subsidence and enhancing the solar radiative heating (e.g., Della-Marta et al. 2007, Beniston and Diaz 2004, Black et al. 2004, Fink et al. 2004, Xoplaki et al. 2003b). However, we know that summertime climate in the Euro-Mediterranean region is dominated by the alternation of extratropical weather regimes and subsidence of tropical-dry air by the descending branch of the Hadley cell (Bolle 2002). Moreover, in terms of pressure and temperature anomalies, there is an anti-phase relationship between the eastern and western Mediterranean (Corte-Real et. al 1995). Therefore, the thesis that extratropical blockings are a dominant mechanism associated with the occurrence of extreme conditions could be questionable and the issue of how far the exceptional heatwaves can be explained by anomalous and persistent blocking conditions remains to be addressed.
A relevant issue concerns the temporal scales. Many authors have analyzed the anomalies observed during the 2003 summer heatwave following different approaches (e.g., Cassou et al. 2005, Ogi et al. 2005, Nakamura et al. 2005, Trigo et al. 2005, Black et al. 2004, Beniston 2004, Beniston and Diaz 2004, Fink et al. 2004, Schär et al. 2004). From these studies it turns out that the anomalous temperature pattern of the heatwave is strongly dependent on the timescale considered for the analysis. While the mean summertime temperature anomaly, covering a vast area centered over France, exceeded 4°C (Fig. 1 in Schär et al. 2004), the largest monthly mean anomaly (of about 5°C) occurred over the Euro-Mediterranean region in June and over southern France, Switzerland and south-west Germany in August (Fig. 1 in Fink et al. 2004). Moreover, a considerable different picture emerges when the August 2003 heatwave is analyzed in sub-monthly timescale: maximum anomalies of about 10°C peaked over north-western France during the first fortnight of August (Fig. 1b in Trigo et al. 2005).

Our challenge was to classify extreme months, to explore the potential links between frequent occurrence of extreme weather conditions over Europe and anomalous large scale patterns and to explore in the climate change signal on extreme climate conditions. To this aim we define a methodology able to identify and characterize the occurrence of these episodes. The method was applied to NCEP dataset and to a set of climate experiments conducted using the SXG coupled model (Gualdi et al., 2006).

**RI1.2. Data and Methodology**

We investigate extreme climate conditions over Europe. These are defined as months during which a large number of extreme days occur. In other words, we study “extreme climate events” and their interannual variability (hereafter, simply “extreme events”). Moreover, our analysis is applied to the summertime months June, July and August separately, to explore the potential intraseasonal variability related to the occurrence of heatwaves (as suggested by Cassou et al. 2005).

Extreme events have been defined starting from daily time series of Tmax, linearly detrended, and covering the whole European domain (30°N-60°N, 10°W-50°E). Those time series are used to build a monthly dataset with information about the number of extreme days registered month per month and year per year, at every grid point of the European domain. Specifically, starting from Tmax series (daily data) we have built monthly series of extreme days (monthly data). The possible range of values for the monthly series is 0-30 (31), representing the number of extreme days registered in a particular month.

An important step at this point is how to define extreme days, since the approach adopted could affect the interpretation of the relative impact of specific events (e.g., Beniston and Stephenson 2004). In our analysis we use a non-parametric method based on the threshold approach (IPCC 2001). A certain day becomes an “extreme-day” when Tmax goes beyond the 75% quantile. Only those days in which Tmax exceeds the score of the 75th percentile are counted to build the series representative of the extreme months. We highlight that the score of the 75th percentile is a function of both the grid point and the calendar day: the period 1958-2004 for NCEP reanalysis and the periods 1961-2000 and 2061-2100 for the SXG coupled model experiments. Those scores are anomalies (expressed in degrees) relative to a grid-point and
calendar-day mean value. Defining extremes when anomalies are exceeding “variable thresholds” is a particularly useful method because allows a comparison between extremes occurring in different regions and at diverse elevations.

The series of monthly values of extreme days, i.e. how many extreme days occur in a month, is then used to study the characteristics of the extreme climatic conditions over a specific region, in our case the Euro-Mediterranean. In order to investigate the characteristics of the variability of the occurrence of extreme events over the Euro-Mediterranean region, we apply an EOF analysis to the monthly values time series of extreme days (EM). The result of this analysis will allow us to identify possible leading modes in the interannual variability of the extreme events.

Since the EOF analysis is used in a frequency domain (the EM fields are spatial distribution of frequency of extreme events), grid points with positive (negative) values of the loadings are characterized by more (less) than average days per month with extreme temperatures. To avoid domain dependent results, varimax rotation is applied (e.g. review article from Richman 1986) and the number of components retained for rotation is determined by the break in the scree plot (Cattell 1966).

After identifying the areas where the extremes maximize, our second goal is to verify whether the occurrence of extreme climate conditions over Europe is associated with specific patterns of the large scale circulation. To this aim we explore the potential link between the occurrence of months with a large number of extreme days and patterns of the large scale circulation and SST anomalies based on correlation maps. Because linear correlations could be greatly affected by outliers, we have estimated correlations based on the Spearman rank method (e.g., Spiegel and Stephens, 1961). Correlations are tested at p<0.05. The 95% is in general accepted as a good level of significance when studying extremes (e.g., Cassou et al. 2005, Haylock and Goodess 2004, Xoplaki et al. 2003b).

**RL1.3. A characterization of extreme months as represented by NCEP reanalysis**

We present the classification of the extreme summertime months, based on the frequency of days with temperatures exceeding a threshold. The analysis is performed by computing the EOF of the EM series over the whole European domain (30°N-60°N; 10°W-50°E). Figure 1 presents the varimax-rotated EOF leading loadings of the EM time series. Three leading modes mainly contribute to the interannual variability of number of days per month with maximum temperatures in the right tail of the statistical distribution (extreme hot days). Shaded areas highlight those regions in which the loadings are higher than an arbitrary threshold (i.e., regions in which the local variance maximizes). The EM occurrence maximizes over three European sub-domains: the north-west Europe (hereafter NW-EU; Fig. 1, left column), the Euro-Mediterranean sector (Med-EU; Figs. 1, middle column) and the Eurasia region (EU-Asia; Figs. 1, right column). The results obtained for the different summer months, June (panel a, b, c), July (d, e, f) and August (g, h, i) indicates that there are small intraseasonal changes in the patterns. For more details see Carril et al. (2007).

Hereafter we describe the interannual variability of the extreme events and its link with the large scale circulation and SST, based on heterogeneous correlation maps. These maps are estimated by correlating the PC associated with the EM variability
over a specific region (i.e., NW-EU, Med-EU, EU-Asia) with diverse atmospheric fields and SST.

**RL1.3.1. NW-EU region**
Extreme events in NW-EU are associated with anomalous blocking patterns. Correlation maps between extreme June conditions over NW-EU and a number of atmospheric parameters are presented in Figure 3 in Carril et al. (2007). Anomalies in SKT displays an area of positive correlations over NW-EU centred in the North Sea. Positive (negative) SST anomalies are found in the central-west North Atlantic (SE of Greenland), while a dipolar structure dominates the variability in the Mediterranean region, with positive (negative) correlations in the west-Mediterranean (Balcans). The correlations with the thermal field are particularly strong in the lower troposphere, with coherent correlations discernible in low-level thickness and significant positive (negative) correlations in NW-EU (over Greece, western Turkey and Siberia). Barotropic anticyclonic anomalies dominate over NW-EU, with significant wind anomalies at low and high levels, and significant negative anomalies in precipitation. A belt of cyclonic anomalies extends (SW-NE) from the coast of Morocco up to the south-eastern Europe, covering the entire Mediterranean basin and then extending into central Russia. Moreover, there are significant SLP anomalies in the Artic region, increasing the meridional gradients in the northern portion of the Scandinavian Peninsula and contributing to low-level anomalous cold advections over north-central Russia.

While the summer marches on, the correlations related to extremes in NW-EU intensify (July is shown in Figure 2). This feature found in both the thermal and the circulation fields. The blocking signature extend south-eastward while intensifying, having the local action centre over north Germany. Southward of the blocking system, high-level cyclonic anomalies dominate over the Mediterranean region, NW-Africa and the tropical North Atlantic with strong high-level wind anomalies encircling a cyclonic vortex over the eastern Mediterranean. The blocking is associated with a local warm and dry troposphere, while cold and wet anomalies take place over the eastern Mediterranean.

In August (Figure 5 in Carril et al. 2007), the blocking is located at 60°N over Norway and the Norwegian Sea. A strong dipolar structure appears at 40°N, having the negative (positive) node over the Mediterranean basin, SE-Europe and Middle-East (eastward of the Caspian Sea) with the strongest correlations at higher levels. Signatures of that pattern are visible in the thermal field, while significant deficit (surplus) of precipitation is observed over the blocking region and in the neighbourhoods of the Caspian and Aral Seas (the Mediterranean basin and south-eastern Europe). At sea level, negative pressure anomalies are still persistent at high latitudes, giving significant westerlies anomalies and probably affecting the intensity and location of the storm tracks.

**RL1.3.2. Med-EU region**
In the early summer extremes over the Med-EU region are associated with barotropic patterns, while in July and August, baroclinic structures became dominant.

Frequent extremes in Med-EU in June (Figure 6 in Carril et al. 2007) are linked with anomalous warming of the Mediterranean SSTs and positive SKT anomalies over
central Europe and NW-Africa, while coastal Atlantic SSTs remain almost unperturbed. The anomalous warming is observed at all tropospheric levels (but larger correlations are at lower levels) and it is associated with barotropic positive anomalies in circulation.

During July (Figure 3), extreme events in Tmax and Tmin are connected with somewhat diverse climatic anomalies. The main difference concerns the anomalous warming of the Mediterranean SSTs. When there are extremes in Tmin the anomalous warming is even more extended: it peaks in the Mediterranean Sea and extends over the full basin and over land coastal points of North Africa and Eastern Europe. On the other hand, extremes in Tmax occur when there is anomalous warming over the western part of the basin and over a large portion of the European continent. In the Tmax case, significant cyclonic SLP anomalies are confined in the neighbourhoods of the Iberian Peninsula. On the contrary, when extremes are in Tmin cyclonic anomalies also extend over north-east Africa, advecting warm air from the Sinai into the eastern Mediterranean sector. There are also noticeable differences in the high-level circulation. Extremes in Tmin are associated with a dipolar structure with positive (negative) nodes centred in the neighbourhoods of Italy (Caspian Sea). The local correlations with the thickness are larger in the Tmax than in the Tmin case, but for the Tmin extremes, atmospheric correlations almost vanish at high-levels (not shown).

The main characteristic of the anomaly circulation associated with extremes over Med-EU in August is the lack of significant barotropic structures (Figure 8 in Carril et al. 2007). At low levels there are generalized cyclonic anomalies, with significant anomalous convergence in the subtropical Northwest Africa, suggesting anomalies in the African monsoon (correlations in precipitation denote the northward displacement of the convective systems; not shown). The surface warming maximizes over the central Mediterranean, but large positive SKT anomalies affect also the central Europe, Balkans and Mediterranean coastal lands in Turkey, Middle-East, North Africa and Morocco. The anomalous heating is also visible in thickness anomalies. The amplitude of the positive node in thickness correlations diminishes with increasing height while the structure of the pattern becomes similar to that in high-level circulation. A notable feature displayed in Fig. 8f-8g in Carril et al. (2007) is the baroclinic sea-saw that dominates the variability in the subtropical region, with opposite nodes in the Pacific and Atlantic sectors (positive streamfunction anomalies in the southern hemisphere are cyclonic anomalies).

**RL1.3.3. EU-Asia**

Extreme events in EU-Asia are associated with anomalous blocking patterns. The intensity of the correlations and the related systems are remarkably modulated by intraseasonal variability.

During June (Figure 9 in Carril et al. 2007), the blocking signature is particularly strong. Positive anomalies in eddy streamfunction are centred over NW-Russia with significant high-level wind anomalies and divergence at surface levels. The thermal field denotes positive anomalies in skin temperature and low-level thickness, and a zonally elongated pattern of positive high-level thickness anomalies that connects the continental anomalies with those upstream of the blocking, associated with northward displacement of the Azores anticyclone.
One month later (Figure 4), the blocking related correlations diminish while the
blocking action centre moves southward. The local warming is also visible in
thickness anomalies. Cyclonic anomalies encircle the blocking structure at low levels,
and significant negative correlations in SLP covers large portion of Europe,
Mediterranean Sea and North Africa. At high levels there is a strong circulation
gradient downstream of the blocking anomalies, with alternant signs and barotropic
structure, which resembles a wave train propagating into Southeast-Asia.

By the end of the summer (August is in Figure 11 in Carril et al. 2007), the anomalous
climatology related to extremes in EU-Asia seems to be organized in a wave pattern
with positive (negative) circulation anomalies affecting the Azores anticyclone
(Iceland low) up to reaching the blocking sector. While anomalies in skin temperature
and low-level thickness are non-significant in the central North Atlantic, the other two
nodes are significant. Downstream of the blocking, the perturbation affects also the
Middle-East and central Asia.

**RL1.4. Discussion**

The results we obtained from the analysis of the extremes in surface temperature over
Europe show well defined anomaly patterns in the general circulation and SSTs.
Extreme events over the north-west Europe and over Eurasia are typically related to
blocking anomalies. The intraseasonal variability of those patterns is related with the
amplitude of the correlations, the relative location of the action centers and the
wavetrains downstream or upstream of the extremes centre. On the other hand,
extremes over the Euro-Mediterranean region are associated with weak blockings in
June and with baroclinic pattern of anomalies in July and August, probably related
with the regional shifts of the descending branch of the Hadley circulation. Extremes
over the Euro-Mediterranean region in July are also linked with the anomalous
warming of the Mediterranean SSTs.

We also notice that large scale circulation anomalies associated with extreme events
sometimes resemble the patterns of interannual variability found in other works. For
example, the Mediterranean dipole in our Figure 2 resembles the anti-phase anomalies
described by Corte-Real et al. (1995) which they have attributed to blocking situations
over the North Atlantic accompanied by depressions in the Mediterranean sector. A
similar picture is also in Figures 5 and 6 by Xoplaki et al. (2003b; hereafter X03b)
were the east-west dipole in the Mediterranean summertime temperature anomalies is
described as associated with a jet stream meandering between cyclonic and
anticyclonic anomalies extending from SE Greenland to the Caspian Sea. It is
interesting that different approaches recover similar patterns. If we refer to extreme
events in the Mediterranean region, the patterns in Figure 6 from Carril et al. (2007)
appear like those in Figure 3 by X03b when describing warm conditions over the
central Mediterranean associated with local anomalous blocking conditions,
subsidence and stability.

Interestingly our patterns present a number of similarities with results from X03b,
although we apply a completely different approach. The main difference is that X03b
studied summertime temperature series, while we consider only temperatures
exceeding thresholds, in order to focus on the variability of extreme cases in the
frequency domain. Therefore, the data considered in the two studies contain different
kind of information and our approach is based on the variability of the portion of spectrum related to the extreme cases. Moreover, in our analysis we have generalized the results over diverse European sub-domains and included the intraseasonal variability.

Furthermore, the similarities between the large-scale anomaly patterns associated with extremes we have found and the patterns of interannual climate variability found in other studies raise the important question as to whether the patterns of interannual variability are biased by the occurrence of extremes. Special attention must be devoted in exploring how much the modes related to the variability of extreme events are different from those explaining the “no-extreme variability” or, in other words, are extremes occurring as a result of a peak in the activity of some particular climate variability mode? To address this point, we have repeated our analysis using time series in which the extremes from the upper and lower tail of the statistical distribution have been removed (no-extreme cases are when temperature is falling in the interval P25 to P75). In this case, the variability spreads into a large number of modes each of one explains few percentage of variance. Modes explaining the variability of the extreme events are also present in the no-extreme case suggesting that extreme events occur when some mode of the “no-extreme variability” strongly peaks. This result seems to suggest that extreme climatic events, as defined in our approach, may occur as a strong peak of a mode of climate variability.

**RL1.6. Modelling extremes and climate change projections**

Although the society in general is vulnerable to the increasing extreme events (Kunkel et al., 1999), Southern Europe is expected to find itself in a situation particularly compromising in a near future, threat from both heatwaves and the reduction of water resources (IPCC 2001). The level of knowledge regarding both the local-interactions and the anomalous climatic conditions in which extremes are embedded is the first constraint for their seasonal prediction, as well as preparing our society to mitigate the effects of those more severe and frequent extremes that are expected to occur in a warmer climate (e.g., Klein Tank and Können 2003).

Whereas in a classical framework the study of heatwaves is done by analyzing singular weather systems that favors local anomalies, weather and climate are sometimes overlapping concepts. There are some years during which those weather systems are more frequent and the successive occurrence of heatwaves is perceived as an extremely anomalous month, with disastrous consequences for humans, ecosystems and lands that are critically stressed without having the necessary time to recover.

Previously, we have described the large scale circulation associated with those extreme months signed by the imprints of heatwaves from NCEP reanalysis. Hereafter, the same methodology to detect extreme climatic conditions and to classify heatwaves over Europe is applied to a coupled model simulation. The analysis is now based on numerical experiments performed using the SXG model (Gualdi et al., 2006). The experiments selected were those of the 20th century climate (SRES 20c3m) and the climate change projection SRES A1B. We have applied the method to 40-year time slices covering the periods 1961-2000 and 2061-2100 respectively.
Both, in NCEP and in the model simulations, we have identified three regions of relevance to study extreme events over Europe: the north-west Europe, the Euro-Mediterranean region and Eurasia. Nevertheless, leading modes accounting for the variability of the extreme days as represented by the present-day climatology of SXG are somewhat different from those obtained from NCEP.

From NCEP,

- Over north-west Europe, extreme events are typically related to blockings. Anomalous barotropic structures are particularly strong (weak) in July (June), when anomalies seems to be organized in a wavetrain patterns that propagates from USA to NW-EU (from the blocking region into Asia). In agreement with results by other authors (Xoplaki et al. 2003a, Xoplaki et al. 2003b, Corte Real et al. 1995), we also find an opposite relation between the pattern of anomalies in the blocking region and over the eastern Mediterranean. Blocking anomalies in August disrupt a particularly perturbed extratropical circulation. The blocking is the counterpart of a dipolar structure that dominates the anomalous variability over Europe with the negative barotropic node over the Euro-Mediterranean sector.

- Over the Euro-Mediterranean region in June, anomalous climate conditions associated with extremes are related with weak blocking situations. In July and August baroclinic patterns of anomalies dominate the variability related to the occurrence of extreme events over that region. In a blocking condition, the physical processes that might produce local intense heating anomalies are the surplus of short wave radiation reaching the surface and the vertical motion favoring the adiabatic warming. Baroclinic anomalies, on the other hand, might be related to regional changes (shifts) of the descending branch of the Hadley circulation.

- Over Eurasia, blocking anomalies dominate the extreme events variability. The large scale anomalies for June mainly differs from that for July in the intensity of the patterns (their amplitude diminishes in July) and in the location of the main cyclonic node, downstream of the blocking. In June (July) a wave train appears to propagate from the blocking pattern into North (South) Asia. Blockings in August appears as the last node in a wave train propagating from the central-west North Atlantic to EU-Asia.

Global climate models provide grid average values. Moreover, models not necessarily provide instantaneous maximum values (every so often, Tmax represents the mean value among instantaneous Tmax values occurring in a period of time of some hours). Then, climate models tend to give lower magnitude of extremes, extended over large areas.

In the present climate, the simulated modes of extreme variability in July are the modes that better compares with those obtained from NCEP. In June and August the model misrepresents the areas of homogeneous variability in extremes, modes that are mainly related with blocking episodes. Blockings occur in simulations, but the model misrepresents their location and intensity.
In particular, Figure 5 presents the extremes related variability as represented by the 20th century experiment in July (compare top, mid and bottom panels in Figure 5 with Figures 4, 2 and 3 respectively). On the one hand, the model captures the baroclinic anomalies associate with the occurrence of extremes over the EU-Med region in July. On the other hand, the model reproduces the extremes over both, NW-EU and EU-Asia region associated with blockings in the extratropical circulation. In these cases, local anomalies in ST, thickness (not shown), circulation and precipitation (not shown) over NW-EU are in agreement with those reported using NCEP reanalysis, nevertheless, anomalies far away from the node does not fully agree with findings from Carril et al. (2007).

The climate change projection illustrated in Figure 6 denotes more intense events in NW-EU, less intense events over EU-Asia, while few changes are observed over the Med-EU sector. Nevertheless, this results must be taken with caution, since we are analyzing changes in extremes related with changes in the circulation, but soil moisture feedbacks are a priori not considered.
Figure 1: Rotated EOF loadings from EM series, displaying the areas of frequent extremes. Isolines are every 3 non-dimensional units, the zero isoline is removed and shading emphasizes the regions where the number of days per month with extremes temperatures has the largest variability. The leading structures denote regions of high variability over three European subdomains: NW-EU (left column), Med-EU (middle column) and EU-Asia (right column). Top-row is for June, middle-row for July and bottom-row for August.

Figure 2: Correlation maps between the PC associated with the variability of extremes in Tmax over the NW-EU in July, and diverse large-scale fields: a) skin temperature, b) high-level eddy streamfunction and winds at 200 hPa, c) low-level thickness, d) sea level pressure and surface winds e) precipitation and f) is an extended view of correlations with high-level eddy streamfunction. Contours are every 0.1 (but every 0.2 for the precipitation), the zero line is omitted and significant anomalies at the 95% confidence level are shaded. Correlation vectors with modulus lower than 0.5 are obscured.
**Figure 3:** Correlation maps between the PC associated with the variability of extremes in Tmax (left column) and in Tmin (right column) over the Med-EU in July, and diverse large-scale fields: a-d) skin temperature, b-e) high-level eddy streamfunction and winds at 200 hPa, and c-f) sea level pressure and surface winds. Contours are every 0.1, the zero line is omitted and significant anomalies at the 95% confidence level are shaded.

**Figure 4:** Correlation maps between the PC associated with the variability of extremes in Tmax over the EU-Asia in July and a) skin temperature, b) high-level eddy streamfunction and winds at 200 hPa, c) low-level thickness, d) sea level pressure and surface winds, e) high-level eddy streamfunction (extended view) and f) low-level thickness. Contours are every 0.1, the zero line is omitted and significant anomalies at the 95% confidence level are shaded. Correlation vectors with modulus lower than 0.5 are obscured.
Figure 5: Correlation maps between the PC associated with the variability of extremes in Tmax and diverse fields (surface temperature in the left column, high-level eddy streamfunction in the mid column and SLP in the right column). Analysis is based on the 20th century experiment using the SXG climate model, for July. Top/mid/bottom panels are for extremes in EU-Med/NW-EU/EU-AS region. Contours are every 0.1, the zero line is omitted and significant anomalies at the 95% confidence level are shaded.

Figure 6: As Figure 5, for the climate change projection.
Research Line 2,

*Relationships between regional and large scale circulation and extreme temperature events in the Eastern Mediterranean.*

Maheras P., K. Tolika, C. Anagnostopoulou, E. Flocas and M. Hatzaki

*Aristotle University of Thessaloniki, AUTH, Greece*

**RL2.1. The role of the regional scale circulation**

Changes in extreme temperature over the Greek area are examined in relation to the geopotential thickness field (1000-500) hPa, for the period 1958-2000 (circulation type classification). Table 1 summarizes their relative frequency of all the circulation types. Extreme temperatures were investigated not only by the mean seasonal maximum and minimum temperature but also by applying two extreme indices: 1) the 90\textsuperscript{th} percentile for maximum temperature (Tmax90p) and 2) the 10\textsuperscript{th} percentile for minimum temperature (Tmin10p).

The analysis of the extreme temperature trends showed that in the case of winter Tmin (Table 2) the study area (Greece) is characterized by a decreasing trend which is statistically significant over the southern Peloponnesus and some Aegean islands. Conversely, a trend of increasing Tmin is apparent during summer. In spring and autumn a statistically significant negative tendency predominates. Generally the number of stations with positive and negative trends is equal but the magnitude of the negative trends is higher than that of the positive ones. Concerning the Tmin10p index it seems that the stations under study present a general decreasing tendency on an annual basis. It should be mentioned that the negative values in spring and autumn mostly account for this annual decreasing tendency (Table 2).

Regarding maximum temperatures, negative trends are found over the Greek area during winter. On the other hand the summer maximum temperatures appear to increase in all stations. For the two other seasons, the sign of the trend seems to vary from station to station with positive (negative) sign being more prominent in spring (autumn). Finally, the annual trend of increasing Tmax90p seems to be mainly associated with the positive values during spring and summer. In contrast, the winter Tmax90p presented negative trend values over the whole area (Table 3).

Aiming on the investigation of the relationship between temperature and circulation types, the deviation of the mean Tmin or Tmax associated with each circulation from its average value (anomalies) for each station were computed on an annual and seasonal basis. Circulation types Ane, A, Asw and AE form strong positive
anomalies, while Anw forms weak positive or negative anomalies. It is worth mentioning at this point that on an annual basis (not shown) the anticyclonic (cyclical) types are characterized by positive (negative) anomalies (Figure 7). Moreover, the winter trends of Tmin anomalies (Table 4) present a decreasing tendency which becomes evident for both anticyclonic and cyclonic types with very few exceptions. This decreasing tendency may be partially attributed to the general increase in frequency of surface anticyclonic types which is related to northerly, northwesterly or northeasterly airflows over Greece. Moreover, it is associated with the increase in frequency and persistence of the surface cold anticyclonic types over the Mediterranean and Balkans.

Respectively, Figure 8 depicts the magnitude of the anomalies of summer Tmax for each circulation type and station. Circulation type A presents the greatest positive anomalies, while the circulation types C, Csw and Cne present the greatest negative anomalies of almost all stations. It is suggested that the increase in frequency of the anticyclonic types can explain the trend of increasing summer Tmax. This leads to an identified westerly to southwesterly wind advecting warm air masses from Northern Africa over Greece or an anomalous northeasterly to easterly continental flow and subsidence at the upper levels.

Moreover, from the study of the maximum and minimum temperature trends, on a monthly basis this time, a significant cooling was shown for November and December over the Greek region. The analysis of the thickness circulation types for these two months indicated that the decrease (increase) in frequency of the anticyclonic (cyclonic) circulation types could contribute to the general decrease of the November/December monthly temperature. An equivalent analysis made for the surface circulation showed an increase in the anticyclonic types (Figure 9), which favor decrease in minimum temperature. The summer warming which is evident for the latter half of the data mainly in June, July, August and September can be explained by decreasing tends in the frequency of the circulation types mostly responsible for the Etesian winds and an increase in the Mb type (slack pressure field, almost absence of a pressure gradient; Figure 10), both leading to more calm days. In addition the positive trend of the frequency of all anticyclonic types during summer lead to increased westerly/southwesterly flow, advecting warm air masses from Northern Africa.

**RL2.2. The role of large scale circulation**

As a significant component of the natural climatic variability, the teleconnections play an important role in changes of the regional climate, and especially of extreme events. In the Mediterranean region, the most prominent influence of independent large scale circulation mode is the North Atlantic Oscillation (Feidas et al. 2004). More specifically, the Mediterranean region is characterized by cooler and drier conditions during the positive phase of NAO, while the negative phase is related to anomalously warm and wet conditions. The role of other teleconnection patterns on Eastern Mediterranean temperature and precipitation regime have also been demonstrated,
such as North Sea Caspian pattern (NCP) (Kutiel et al., 2002) and the Eastern Atlantic-West Russia Pattern (EAWR) (Krichak and Alpert, 2005).

In this study, the impact of an upper level teleconnection pattern, called Eastern Mediterranean Pattern (EMP) on the extreme temperature regime of Eastern Mediterranean is investigated in the present and future climate. EMP was identified between Eastern Mediterranean and Northwestern Europe in the 300 and 500 hPa geopotential field during winter (Hatzaki et al., 2007). The negative and positive phase of the EMP were discriminated and it was found that during the negative phase a positive anomaly of 500hPa geopotential heights forms over the Eastern Mediterranean and northern Africa, implying an increased anticyclonic circulation, while during the positive phase a cyclonic anomaly predominates.

The following datasets were employed for the present period 1958-2000: a) daily geopotential height, being obtained from the NCEP/NCAR Reanalysis Project for the isobaric level of 500hPa on a 2.5°x2.5° grid and b) daily station data of maximum and minimum surface temperature data for 26 stations (Figure 11). For the future climate, daily datasets of the 500hPa geopotential height derived from the Global Circulation Model HadAM3P on a 2.5°x2.5° grid and of maximum and minimum temperature derived from the Regional Climatic Model HadRM3P with a resolution of 0.44° x 0.44° for the period 2070-2100, using the IPCC emission scenario B2 for the evolvement of the future atmospheric concentrations of greenhouse gases.

The investigation of the impact is based on the Regularized Canonical Correlation Analysis (RCCA) while qualitative estimations of the composite temperature anomalies are performed for each phase of EMP. The positive and negative phases of EMP were identified with the aid of the corresponding standardized index, following the methodology described in Hatzaki et al. (2007). More specifically, the index was defined in the present climate as follows: \( EMPI = gpm(25°W, 52.5°N) - gpm(22.5°E, 32.5°N) \), where gpm is the mean winter geopotential height (in gpm) of the grid point which form each pole, respectively.

For the future climate, Hatzaki et al. (2006) have demonstrated that the overall pattern is retained; however, the two EMP poles appear to be shifted in relation to their present positions. In particular, the whole pattern is prominently moved to the south during the B2 scenario while the shift is greater for the northern pole. Therefore, due to apparent shift of the two poles of the pattern, the EMPI needs to be revised, as compared to the definition for the present climate. The index is then re-defined for B2 scenario, as follows: \( EMPI_{B2a} = gpm(2.5°W, 47.5°N) - gpm(32.5°E, 30°N) \).

The identification of the climatic extremes was performed with the aid of the following climatic indices that were employed within the context of STARDEX a) Tmax90 (90\(^{th}\) percentile of maximum temperature), representing extremely warm episodes b) Tmin10 (10\(^{th}\) percentile of minimum temperature), representing extremely cold episodes.
The analysis of the canonical loadings of the second CCA pair for maximum temperatures, with correlation 0.97, showed that positive (negative) geopotential height anomalies lead to increased (decreased) maximum temperatures, mainly over the central and northern examined region (not shown). The same analysis for the minimum temperature showed that the EMP is not depicted in the 500 hPa geopotential heights field, thus its impact on Tmin could not be extracted (not shown).

According to Figure 12, where the composite anomalies of the indices Tmax90 and Tmin10 are displayed, a decrease becomes evident during the positive phase for both indices over the examined area. The statistical significance of the anomalies has been examined at the 0.05 level. This decrease is more prominent for Tmin10, where a peak of -0.9 °C is found over the Aegean Sea and of -1.1°C over northern Greece, suggesting an increase of the extremely cold episodes. The opposite behavior is observed during the negative phase, when both Tmax90 and Tmin10 increase, by 1°C and 1.3°C, respectively, especially over continental Greece.

Under the B2 scenario, the anomalies of the index Tmax90 during the two phases of EMP follow a similar distribution to the present climate with negative (positive) biases during the positive (negative) phase, apart from western Greece which has the opposite behavior from the rest of the examined area (Figure 13). Concerning the Tmin10, the future positive phase is characterized by an overall decrease in the entire examined area, similar to the corresponding present phase. During the negative phase, an apparent increase is found (Figure 14).

It was found that the EMP affects indeed the extreme events with inverse impacts between the two phases. The highest anomalies are found over continental Greece. More specifically, a positive phase of EMP is associated with a temperature decrease while the opposite occurs for the negative phase. In the future, the present impact according to each phase persists and intensifies, due to the estimated future shift of the EMP poles.
Table 1: Seasonal and annual frequencies (%) of the 14 circulation types for the period 1958-2000. Acronyms are circulation types: A) anticyclonic and C) cyclonic, followed by location of positive or negative anomaly centres in relation to Greece (absence of location information indicated anomaly centre directly over Greece).

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Table 2: Trends of mean seasonal and annual minima: a) Daily minimum temperatures (Tmin) and b) 10th percentile for minimum temperature (Tmin10p). Statistically significant boxes are shaded.
Table 3: Trends of mean seasonal and annual maxima: a) Daily maximum temperatures (Tmin) and b) 90th percentile for maximum temperature (Tmax90p).
Statistically significant boxes are shaded.

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Table 4: Trends (positive or negative) of winter Tmin for each circulation type. Statistically significant boxes are shaded. Zero means no trend.
Figure 7: Winter minimum temperature anomalies (°C) at the 20 Greek stations for the 14 circulation types (1-14 for the acronyms see Table 1).

Figure 8: As in Figure 7, but for maximum summer temperature anomalies.
**Figure 9:** Frequencies of the six SLP anticyclonic types in the case of December (1958-200).

**Figure 10:** Frequencies of the Mb circulation type for the period June-September (1958-200).
**Figure 11:** Geographical chart of the Eastern Mediterranean Basin, where the location of the employed stations is displayed.

**Figure 12:** Biases of winter mean value of the indices $T_{max90}$ and $T_{min10}$ during the a) positive and b) negative phase of EMP for the present climate (1958-2000). The dashed contours indicate negative values.
Figure 13: Biases of winter mean values of Tmax90 during the positive and negative phase of EMP for the future climate (2070-2100). The dashed contours indicate negative values.

Figure 14: As in Figure 13, but for Tmin10.
The increasing risk of extremely long melt seasons in the Arctic due to global warming

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RL3.1. Introduction

The Arctic has undergone a warming the past 100 years that is twice the global observed warming, this temperature change being consistent with changes in observed cryosphere and oceans (IPCC WG1 FAR 2007; ACIA 2005). In this paper we will investigate trends and variability in the length of the Arctic melting season. A lengthening of the melt season may have a number of consequences. It would lead to a further thinning of the Arctic sea ice (Laxon et al. 2003), reduction of the permafrost (IPCC WG2 FAR 2007; Hinzman et al. 2005; ACIA 2005) and a lengthening of the growing season (Smith et al. 2004). With these expected and on-going changes the whole appearance of the Arctic will be altered. Permafrost for instance, is recognized as a solid ground for constructions, but warming will make the constructions and infrastructure more vulnerable (IPCC WG2 2001; ACIA 2005). Moreover, the lengthened growing season would lead to changes in vegetation and import of new species, this putting an increased stress on existing ecology. Actually, the growing season is reported to have increased with 1-3 days per decade depending on the region in the Arctic (IPCC WG2 FAR 2007). In April and May 2006 record high temperatures were recorded in the Barents Sea region. The media recorded that ice roads at Spitsbergen north in the Barents Sea were useless earlier than normal, and summer season started earlier than ever recorded creating a very long melting season.

In addition to the analysis of the Arctic melting season, we also explore causalities in a statistical. Furthermore, we estimate the probability of a very long melting season within the present climate, and in projected climate.

RL3.2. Data

The homogenised time series of observed monthly mean surface temperature for the period 1912 to 2006 at Svalbard Airport, Spitsbergen (Nordli and Kohler 2004) is used. This time series is a composite of several shorter series of measurements at a few nearby sites. All the shorter time series are adjusted (Nordli and Kohler 2004) to
be valid for the current Svalbard Airport weather station (78°15′N, 15°28′E, 28 m. a.s.l) established in 1975. The daily time series from 1975 to 2006 is also compared with daily temperatures estimated from monthly data (see below).

The global monthly mean temperature anomaly of Northern Hemisphere (from the 1961-1990 mean) will later be used as a predictor in a linear model (HadCRUT3v) is downloaded from the website http://www.cru.uea.ac.uk/cru/data/temperature/ and is described in (Jones et al. 1999). The dataset is one of many possible datasets of global temperature, but the results do not depend much on the choice of global temperature dataset (IPCC WG1 FAR 2007). Sea Level Pressure (SLP) data is taken from HadSLP2r dataset (Allan and Ansell 2006). The NAO index is the difference between the normalised sea level pressure over Gibraltar and the normalised sea level pressure over Southwest Iceland (Jones et al. 1997).

RL3.3. Impact of melting season

To quantify the impact of the length of the melt season, \( \tau \), it is convenient to split the annual mean temperature into a contribution from the cold season and a contribution from the melt season:

\[
\overline{T}_s = \frac{(365 - \tau)}{365} \overline{T}_c + \frac{\tau}{365} \overline{T}_m
\]

Here, \( \overline{T}_s \) is the annual mean temperature at Svalbard Lufthavn, \( \overline{T}_c \) is the mean temperature during the melting period \( \tau \), and \( \overline{T}_m \) is the mean temperature during the rest of the year. The melt season \( \tau \), is defined as the number of days with surface temperature above 0°C. We also include the requirement that there should be a cluster of 4 or more consecutive days that exceeds this threshold in order to contribute to \( \tau \) (\( T_{\text{daily}} > 0°C \)). According to IPCC WG1 FAR (2007) the sea ice border has around 100 days duration of melt season while the central Arctic has approximately 50 days. Spitsbergen located close to the sea ice border north in the Barents Sea is then a very typical place to investigate the changes in melt season. The border of the Arctic ice sheet is the area with expected largest changes in melt season in the future according to IPCC WG1 FAR (2007).

Records of daily time series of temperature in the Arctic are short and often with missing data. To overcome this problem, when calculating \( \tau \), we have employed the smoothing spline method by de Boor (1978), to reconstruct daily measurements from monthly data. A method based on harmonic fit has also been introduced for this purpose Epstein (1991). However, both methods will fail in reproducing extreme daily event and high day-to-day variability since the method is based on passing a smooth curve through the monthly values providing climatology of daily values. Smoothing spline is preferred, because of it accuracy and simplicity (de Boor 1978). The smoothing spline method is compared with the daily observations for 2006 in Figure 15. The daily observations are as expected noisier than the estimated daily climatology. The parameters in the smoothing splines method is estimated by minimizing the maximum likelihood of the difference between the observed with the reconstructed \( \tau \).
The reconstructed and observed $\tau$ at Svalbard Airport with $T_{4,\text{daily}} > 0^\circ\text{C}$ is shown in Figure 16a. The mean melt season duration for the period 1912 to 2006 is estimated to be 105 days and the growth rate is $1.5 \pm 0.4$ days per decade. According to IPCC WG1 FAR (2007) the sea ice border has around 100 days duration of melt season while the central Arctic have approximately 50 days. Spitsbergen is located in the marginal ice zone. The estimated trend is quite similar to the growth season change over the same period (IPCC WG1 FAR 2007), which is between 1 to 3 days per decade depending on location in Arctic. A paired two-sample t-test with removed linear trend shows that there are no significant differences in mean between the observed in mean melt season and estimated melt season from smoothing spline. The linear correlation is estimated to be $r = 0.82 \pm 0.09$ between observed and estimated melting season. The estimated daily temperature fits well with the observed number of days with $T_{4,\text{daily}} > 0^\circ\text{C}$ with a few exceptions, but the annual fluctuations are caught up by the estimated time series.

For the period 1976-2006 the growth rate has been $8.7 \pm 2.3$ days per decade. The differences between the first day estimated from monthly mean data and observed daily observation is caused by short lived dynamical system. This variability is expected since the estimation of daily from monthly data cannot catch such short lived systems. There is a significant trend for both the first and last day, the first day of $T_{4,\text{daily}} > 0^\circ\text{C}$ occur earlier and the last day of $T_{4,\text{daily}} > 0^\circ\text{C}$ later in the year (Figure 16b).

But what is the most important factor contributing to Arctic amplification, increased temperature or increased melt season length? The mean annual temperature ($\overline{T}_a$) is plotted in Figure 17a. A simple regression shows a significant trend of $0.2^\circ \pm 0.12^\circ\text{C}$ per decade for the mean temperature. The significance level is $\alpha = 0.05$. The simple equation 1 is used to figure out which component $\overline{c}_T$, $\overline{m}_T$ or $\tau$ contribute most to the warming trend. $\overline{c}_T$, $\overline{m}_T$ and $\tau$ have a significant positive trend with $\alpha = 0.05$. However, the p-value is lowest for $\tau$, and this is the most important factor for the change in $\overline{T}_a$. $\overline{c}_T$ and $\overline{m}_T$ are plotted in Figure 17b-c, and there is no significant differences between the observed and reconstructed $\overline{c}_T$ and $\overline{m}_T$.

In order to obtain indications of which processes that are involved in the changes in the melt season at Svalbard, we employ a linear regression model where $\tau$ is linearly related to a set of predictors that represent different processes in the climate system. One obvious candidate is global warming. The annual global mean (or Northern Hemisphere mean) temperature can thus be used as a predictor in our model. This predictor could be said to represent the slowly evolving thermodynamic changes due to the increased greenhouse gas forcing. A second class of predictors can be associated with flow changes. Following Hartmann (1994), it is the synoptic part of the atmospheric flow that dominates the pole ward heat transport in the Northern Hemisphere summer. According to Seierstad et al. (2007) there is evidence from modelling and observational studies that there exist links between the large scale and synoptic part of the flow. Seierstad et al. (2007) demonstrated this by means of a Generalised Linear Model. They showed that the local large scale flow (represented by the monthly sea level pressure) can account for up to 30% of the variance in the
synoptic flow (or storminess). Adding teleconnection indices in the predictor list gave added explained variance, but in limited geographical regions at high latitudes. This, and related work (Seierstad et al. (2007) and references therein) is based on winter-time data. To our knowledge very few other studies have explored summer-time synoptic activity, except for Mesquita (2006) and Serreze et al. (2000). In Mesquita (2006) it is shown that the following parameters could be statistically linked to the synoptic flow variations: North Atlantic Oscillation (NAO), East Atlantic jet (EAjet), West Pacific Pattern (WP) and Atlantic Multidecadal Oscillation (AMO). However, only NAO index is available for our time period of time, and included in our model. The pressure variability is also included in the model.

Further, surface ice and surface temperature are highly correlated. According to Belchansky et al. (2004), there were a high correlation between the winter Arctic Oscillation (AO) and sea ice melt duration the following summer. Since AO and NAO are highly correlated (Ambaum et al. 2001), the winter and fall NAO is used to describe possible variations in Arctic sea ice melting.

There is a trend over the last century towards longer $\tau$. We think it is likely that this is a local manifestation of anthropogenic global warming. According to IPCC WG1 FAR (2007), the global temperature has risen about $0.74 \pm 0.18^\circ C$ the last century. On the other hand, we see that melt season is affected by short scale fluctuations. We thus propose a linear regression model to investigate the link between the global mean temperature anomaly and length of melt season. The length of melt season also depends on thermodynamic processes that can be related to global mean warming and more short living dynamical factors such as spring and autumn heat wave events. By using Sea Level Pressure (SLP) data taken from HadSLP2r dataset (Allan and Ansell 2006), two components of pressure difference between Iceland and Kara Sea to describe the advection of warm air from the South for the spring and the fall are introduced. These indices are standardized, and this pressure difference has a very similar spatial pattern with the Barents Oscillation (Skeie 2000). The regional summer variation of SLP for the area of Svalbard is also included in the model.

In table 5 several potential regression models have been tested. The model 7 explain most of the variation ($R^2 = 0.2812$). However, only $T_{\text{ice}}$ and $p_{\text{BO}}$ are statistically significant. The components suggested in Belchansky et al. (2004) do not have the same influence on $\tau$ as in other regions in Arctic. Svalbard is north of the region investigated by Mesquita (2006), and the NAO index does not explain the variation in $\tau$ in this region. However, the BO index introduced in Skeie (2000) does have some influence on the melt season. This difference describes the temperature advection of warm air into the Svalbard region.

A simple prediction plot is set up in Figure 18 using the parameters from equation 7 in table 5. The temperature increase during this century due to global warming is estimated to be $1^\circ - 6^\circ C$ depending on emission scenario (IPCC WG1 FAR 2007). Our prediction plot shows that at already $3^\circ C$, the melt season can be longer than 6 months. It can come earlier if warm periods above $0^\circ C$ occur in spring and/or fall. It is too few data to make a significant statement on trends of early or late heat events in the warm season, but it is worth to note that two of three long lasting anti-cyclone events causing melting are observed in the latest decade. The linear model is compared with the non-linear model lowess smoother that make use of locally-
weighted polynomial regression (Cleveland 1979, 1981) for different smoother parameters. The non-linear models are within the prediction level. However, the difference between the linear and non-linear models can be explained with the extremely long melt season influence on the trend estimation.

**RI3.4. Risk of extremely long melt season**

As mentioned in previous section, the length of melting season has a variety of impacts and it is of interest to figure how the probability of having a melt season longer than 180 days (6 months). A longer melt season has impacts on the decrease of sea ice (Laxon et al. 2003) and land ice (IPCC WG1 FAR 2007). In present climate, the occurrence of melt season of such length can only be obtained by having heat wave events both in spring and autumn. The models presented in table 5 show that the global temperature explains the trend in melt season length, and therefore should be included as a covariate in a model describing the extreme long seasons.

Extreme Value Theory (EVT) is used to investigate the extreme occurrences of melt season and to produce return periods linked to global warming. Values above high thresholds can be studied using generalized Pareto distribution (GPD) (Beirlant et al. 2004; Coles 2001). The threshold approach has been used earlier in climate (Naveau et al. 2005; Nogaj et al. 2006; Coelho et al. 2007). Common for all these papers is that the main focus has been on tropical and/or extra-tropical regions. This is also known as the peak-over-threshold approach for analysing extremes. The so-called block design (Coles 2001) approach cannot be used when the data has only one observation annually. A discussion of the block design versus peak over threshold method is given in Coelho et al. (2007). The probability formula for GPD distribution is defined as:

$$\Pr(X > x \mid X > u) = \left\{ 1 + \frac{\xi}{\sigma} \left( \frac{x - u}{\sigma} \right) \right\}^{-\frac{1}{\xi}} \quad (2)$$

where $x$ is the observations, $u$ is the threshold (also called location parameter), $\xi$ is the shape parameter and $\sigma$ is the scale parameter. The scale $\sigma$ parameter provides an estimate of the variability of the excesses. High values of $\sigma$ have a higher variability of extremes. The shape parameter $\xi$ provides information about the shape of the tail of the distribution of excesses. With covariates it is possible to identify factors that have an influence on shape and scale parameters (Coelho et al. 2007; Coles 2001). For more introduction details about general EVT analysis, see Coles (2001).

One of the difficulties with the peak-over-threshold approach is the right choice of threshold. Too low threshold will violate the asymptotic basis of the model leading to bias, while a too high threshold will generate few excesses leading to high variance. In Coles (2001) p. 78-80 and p. 83 two alternative methods for estimating suitable threshold are given. Both methods indicate that the threshold should not be higher than 115 days and not lower than 110 days (not shown). Since 115 days are also close to the 90th percentile suggested in the IPCC WG1 (2001) as a threshold for extreme events, this threshold is chosen.
Several EVD models with $T_{sah}$, $p_{mno}$, $P_{mean}$ and $p_{BO}$ as covariates are tested. A simple model is tested upon more complex model by using likelihood ratio test (Coelho et al. 2007). Since a more complex model can describe the behaviour of extremes better than, several models have to be tested. The test shows that the more complex models do not explain more significant of the behaviour of the extremes using additional covariates than $T_{sah}$. Both linear and exponential models are tested. The shape and scale parameters are estimated by having the global temperature anomaly of northern hemisphere as a covariate. The parameters are estimated by using maximum likelihood method. The simple model is defined to be:

$$\sigma = \sigma_0 + \sigma |T_{sah}|$$

(3)

$$\xi = \xi_0$$

(4)

where $\sigma$ and $\xi$ is estimated shape and scale parameter in equation 2 by the maximum likelihood method. The threshold is set to be 115. The parameters are estimated to be $\sigma_0 = 4.15 \pm 1.40$, $\sigma_1 = 8.91 \pm 6.33$ and $\xi_0 = 0.25 \pm 0.27$. The shape parameter $\xi_0$ is not significant higher than 0, resulting in that there is a possibility of having an upper limit of melt season length.

The return period is the frequency with which one would expect, on average, a given value (e.g. an excess $z$ of 10 days) to recur. Return period estimates are calculated based on the formula $(1 - H(z))^{-1}$ where $H(z)$ is the probability of having an excess $z$. The return period of an event with duration is given in Figure 19. A 95% confidence interval is calculated by estimating the variance of return period by the delta method (Coles 2001, page 33). The confidence interval is given in grey shades. With the present climate ($T_{sah} = 0$ for the period 1961-1990), the probability of having a melt season with duration 180 days is estimated to be 0.002. If the global mean temperature increase with 6°C, the probability increase to 0.55, more often than every second year. However, with only a global mean temperature increase of 2°C, the chance of having a melting season with duration of 180 days has increased to 0.108. This is approximately every tenth year. A 2°C global mean temperature increase is within the estimates of IPCC WG1 FAR (2007) for present century.

**RL3.5. Discussion and conclusion**

There are short lived dynamical weather systems causing high positive temperature anomalies bringing the $T_{daily} > 0$°C that the smoothing spline cannot catch up. This is partly avoided by exclude clusters with length smaller or equal to 3 days. In other words, short pre- or post-melting periods of the melt season is avoided.

Arctic is a small synoptic system, but with large region differences. The major problem is to locate long lasting time series of temperature, and then the quality of the measurements. However, we have looked at 2 other stations in the same region for a shorter time span. Bear Island (74°31′N, 19°01′E, 16 m. a.s.l, 1923-2006) and Jan Mayen (70°56′N, 8°40′W, 10 m. a.s.l, 1921-2006) are stations located in the Barents’ Sea south of Svalbard Airport. The time series are freely available from the
Norwegian Met Department (http://eklima.met.no). Analysis of melt season records of Jan Mayen and Bear Island show that the return period of melt season events of duration 270 and 180 respectively will decrease with global warming (not shown). The analysis should be expanded for gridded datasets in future studies.

Global and regional climate models can improve the predictions of the future melt seasons. The linear relationship can change in the future, making extrapolation an issue. Simple tests with Bergen Climate Model (Furevik et al. 2003) show that the increase will be linear throughout the next century with the A2 emission scenario (Nakicenovic et al. 2000). The relationship between melting season and global mean temperature anomaly can change in the future, and this should be investigated in future studies. The study should also be expanded for more stations and global gridded datasets, and possibly have a model of global atmospheric CO2 mean content instead of global mean temperature.

Global warming explains mean changes in the length of melt season. Extremely long melt season today is caused by heat waves early or late in the warm season. By using splines, the melt season is estimated from monthly mean temperature. Splines do not catch up shorter periods $T_{\text{daily}} > 0^\circ \text{C}$. The melting season is predicted to increase in length in the future. With global warming, the chance of having a melting season equal to 180 days will change dramatically. The chance of having a melt season with duration 6 months or longer will change from 0.002 in present climate to 0.108 with only 2°C temperature increase something predicted to happen in this region (IPCCWG1 FAR 2007). This study shows with a simple method that a small global temperature increase will have very large influence on melting season in the Arctic. If there is an occurrence of a heat wave both in spring and fall, there is a possibility to have a long melting season as 180 days with the present climate. But with a global warming, this risk will increase dramatically. Extremely long melting season will result in less sea ice, and policymakers should take this into account, when the future strategies for the Arctic economy, ecology and industry shall be decided.
Figure 15: Daily development of temperature in year 2006. The estimated daily climatology is given in gray using smoothing splines (de Boor 1978), while the filled black line indicate the monthly mean temperature.

<table>
<thead>
<tr>
<th>Model no.</th>
<th>Regression Model</th>
<th>$R^2$</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$\tau = \beta_0 + \beta_2 T_{\text{ghi}} + \epsilon$</td>
<td>0.1966</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>2</td>
<td>$\tau = I_1 + \beta_3 p_{\text{NAO}_{\text{AMJJAS}}} + \epsilon$</td>
<td>0.2025</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>3</td>
<td>$\tau = I_1 + \beta_3 p_{\text{NAO}<em>{\text{AMJJAS}}} + \beta_4 P</em>{\text{mean}} + \epsilon$</td>
<td>0.2242</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>4</td>
<td>$\tau = I_1 + I_2 + \beta_5 p_{\text{NAO}_{\text{JMFM}}} + \epsilon$</td>
<td>0.2258</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>5</td>
<td>$\tau = I_1 + I_2 + \beta_5 p_{\text{NAO}<em>{\text{JMFM}}} + \beta_6 p</em>{\text{NAO}_{\text{AMJJAS}}} + \epsilon$</td>
<td>0.2306</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>6</td>
<td>$\tau = I_1 + I_2 + I_3 + \beta_5 p_{\text{BO}_{\text{spr}}} + \epsilon$</td>
<td>0.2630</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>7</td>
<td>$\tau = I_1 + I_2 + I_3 + \beta_6 p_{\text{BO}<em>{\text{spr}}} + \beta_7 p</em>{\text{BO}_{\text{fal}}} + \epsilon$</td>
<td>0.2812</td>
<td>&lt; 0.001</td>
</tr>
</tbody>
</table>

Table 5: The melting season length models for Svalbard Airport, Spitsbergen. The abbreviations are defined as follows: $I_1 = \beta_0 + \beta_1 T_{\text{ghi}}$, $I_2 = \beta_1 p_{\text{NAO}_{\text{AMJJAS}}} + \beta_4 P_{\text{mean}}$ and $I_3 = \beta_5 p_{\text{BO}_{\text{spr}}} + \beta_6 p_{\text{BO}_{\text{spr}}}$. The explanations of the variables are: $T_{\text{ghi}}$ is global mean temperature for Northern Hemisphere, $p_{\text{NAO}_{\text{AMJJAS}}}$ is the NAO index in April to September, $P_{\text{mean}}$ is the sea level pressure mean, $p_{\text{NAO}_{\text{JMFM}}}$ is the winter NAO index, $p_{\text{NAO}_{\text{AMJJAS}}}$ is the fall NAO index, $p_{\text{BO}_{\text{spr}}}$ is the pressure difference (BO index) index between Iceland and Kara Sea in the spring and $p_{\text{BO}_{\text{fal}}}$ is the pressure difference (BO index) index between Iceland and Kara Sea in the fall.
Figure 16: Time series of (a) annual number of days with $T_{\text{daily}}>0^\circ$C estimated from monthly temperature records from Svalbard Airport, Spitsbergen. The red line is the observed measurements using daily dataset from Svalbard Airport for the period 1976 to 2006. Monthly temperature profile (b) of Svalbard Airport displays seasonal variation.
Figure 17: The time series of $\bar{T}_s$ (a), $\bar{T}_m$ (b) and $\bar{T}_c$ (c) for Svalbard Airport, Spitsbergen in the period 1912 to 2006.
Figure 18: Plot of prediction (gray shade) of the linear regression model presented in equation 1.

Figure 19: Return period of a melting season with duration 180 days assuming no change in relationship between melting season and global temperature anomaly.
Research Line 4,

Linking local warm extremes in summer to large-scale circulation structures; validation of ECHAM5/OMI model simulations and evaluation of the SRES A1b scenario

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RL4.1. Introduction

Weather extremes such as extreme wind speeds, extreme precipitation or extreme warm or cold conditions are experienced locally. They are usually connected to circulation structures of much larger scale in the atmosphere. For example, if we restrict ourselves to the Netherlands, a well-known circulation structure that often leads to extreme hot summer days is a high pressure system that blocks the inflow of cooler maritime air masses. Moreover, the subsidence of air in its interior leads to clear skies and an abundance of sunshine that leads to high temperatures. If the blocking high persists and depletes the soil moisture due to lack of precipitation and increased evaporation, temperatures tend to soar, as it did in the European summer of 2003 Schär et al. (2004). Speculations about a positive feedback of dry soil on the persistence of the blocking high can also be found in the literature (Ferranti and Viterbo, 2006).

In order for climate models to correctly simulate the probability of extreme hot summer days, a crucial ingredient is the correct simulation of the probability of the occurrence of blocking. This is a well-known difficult feature of the atmospheric circulation to simulate realistically (Pelly and Hoskins, 2003). The verification of models with respect to this aspect is, in practice, difficult as well, since idealized model experiments suggest a high degree of internal variability of blocking frequencies even on decadal timescales (Liu and Opsteegh, 1995).

In a world with increasing concentrations of greenhouse gases, not only the temperature increases, also the large-scale circulation adjusts to achieve a new (thermo) dynamical balance. Models disagree on the magnitude and even the direction of this change locally (van Ulden and van Oldenborgh, 2006). For instance, a change in the probability of European blocking conditions in summer immediately impacts the future probability of European heat waves. This makes probability estimates of future European heat waves uncertain.

To address the questions concerning the probability of future extreme weather events, and the evaluation of climate model simulations in this respect, it is necessary to have a descriptive method that links local weather extremes to large-scale circulation features. To the best of our knowledge, an optimal method to do so does not exist in the literature. We identified two approaches in the literature to link local weather
extremes to large-scale circulation features. In the first one, the circulation anomalies are classified first, the connection with local extremes is analyzed in second instance. The “Grosswetterlagen” developed by synoptic meteorologists for instance is one such classification Kysely (2002). All kinds of clustering algorithms are another example (Plaut and Simonnet, 2001); Cassou et al., 2005). In our opinion, this approach is not optimal since in the definition of the patterns, information about the extreme is not taken into account.

In the second approach, a measure of the local extreme does enter the definition of the large-scale circulation patterns. For instance, only atmospheric states are considered for which the local extreme occurs. Next a simple averaging operator is applied [“composite method” as in Schaeffer et al. (2005)] or a clustering analysis is performed (Sanchez-Gomez and Terray, 2005). The composite method falls short since it finds by definition only one typical circulation anomaly and from synoptic experience we know that often different kind of circulation anomalies lead to a similar local weather extreme. The clustering analysis is debatable since the data record is often too short to identify clusters with enough statistical confidence (Hsu and Zwiers, 2001; Stephenson and O’Neill, 2004; Berner and Branstator, 2007).

In this report we employ a new optimal method to relate local weather extremes to characteristic circulation patterns. This method objectively identifies, in a robust manner, the different circulation patterns that favour the occurrence of large local weather anomalies. The method is inspired by the Optimal Autocorrelation Functions of Selten et al. (1999). It is based on considering linear combinations of the dominant Empirical Orthogonal Functions that maximize a suitable statistical quantity. The method is described in Panja and Selten (2007) and will be applied in this report to link large local daily temperature anomalies in several locations throughout Europe to anomalies in the 500 hPa streamfunction field in the summer months July and August. The same analysis is applied to observational (ERA40 reanalysis dataset) data as well as climate model data (ECHAM5/OMI) to (1) validate the model simulations and (2) assess the contribution of changes in the circulation to changes in the local temperature extremes.

**RL4.2. Model and Data description**

We analyse an ensemble of climate simulations of the ECHAM5/MPI-OM coupled climate model developed at the Max-Planck-Institute for Meteorology in Hamburg. The model was chosen because it has a realistic meridional overturning circulation in the Atlantic [Marlsand et al., [2003] and performed well on a number of criteria during an intercomparation of all AR4 models, such as the atmospheric circulation over Europe (Van Ulden and Van Oldenborgh, 2005). The two component models, ECHAM5 for the atmosphere and MPI-OM for the ocean, are well documented (ECHAM5: Roeckner et al. [2003], MPI-OM: Marlsand et al. [2003]), and a Special Section of the Journal of Climate was devoted to the coupled model (vol. 19(16), pp 3769-3987). Therefore, we here only give a very short description of the model.

The version used here is the same that has been used for climate scenario runs in preparation of AR4. ECHAM5 [Roeckner et al., 2003] is run at a horizontal resolution of T63 and 31 vertical hybrid levels with the top level at 10 hPa. The ocean model MPI-OM [Marsland et al., 2003] is a primitive equation z-coordinate model. It
employs a bipolar orthogonal spherical coordinate system in which the north and south poles are placed over Greenland and West Antarctica, respectively, to avoid the singularities at the poles. The resolution is highest, O(20-40 km), in the deep water formation regions of the Labrador Sea, Greenland Sea, and Weddell Sea. Along the equator the meridional resolution is about 0.5. There are 40 vertical layers with a thickness ranging from 10 m near the surface to 600 m near the bottom.

The baseline experimental period is 1950-2100. For the historical part of this period (1950-2000) the concentrations of greenhouse gases (GHG) and tropospheric sulfate aerosols are specified from observations, while for the future part (2001-2100) they follow the SRES A1b scenario. Stratospheric aerosols from volcanic eruptions are not taken into account, and the solar constant is fixed. The ensemble runs are initialized from a long run in which historical GHG concentrations have been used until 1950. Different ensemble members are generated by disturbing the initial state of the atmosphere. Gaussian noise with amplitude of 0.1 K is added to the initial temperature field. The initial ocean state is not perturbed. An ensemble of 17 members is generated in this way.

For the analysis of this report, we consider daily time series of 500 hPa streamfunction as an indicator of the state the large-scale circulation and daily mean surface air temperature values over the Atlantic-European area during the months June and July. Two periods are considered; 1958-2000 for validation and 2050-2100 to evaluate the simulated climate change. The model data are validated against the same fields of the ERA40 reanalysis dataset.

**RL4.3. Evaluation of the 1958-2000 period**

**RL4.3.1 Mean and standard deviation**
First we assess the ability of the climate model to reproduce the basic characteristics of the summer circulation by comparing the mean and standard deviation of 500 hPa streamfunction and surface air temperature in Figure 20. The two troughs that are present in the mean circulation are rather well reproduced as well as the mean gradients. The structure of the daily standard deviation is also fairly well captured. The model simulated maximum variability is about 10% higher than observed. The mean temperature distribution is also reasonably well simulated, with the largest discrepancies over the North-eastern part of the North-Atlantic basin and North-Eastern Europe where the model is several degrees too cold. Also the temperature variations are reasonably well simulated with 1-2 degrees over the ocean and 3 to 4 degrees over the continental area with increasing amplitudes moving into the continent.

**RL4.3.2 Warm extremes in the Netherlands**
The Extreme Associated Functions (EAF) methodology of Panja and Selten (2007) is applied to the July-August daily time series of temperature in the Netherlands and the streamfunction field at 500 hPa. For this purpose, the time series are detrended and the seasonal cycle is subtracted. We checked that details of how this is accomplished do not affect the reported results significantly. We focus on the first EAF as this pattern has the largest influence on the local temperature extremes (Panja and Selten, 2007). Figure 21 shows the dominant EAF for both ERA40 (left) and the model. Both
patterns are very similar and correspond to a blocking high centered northeast of the location of the local temperature time series (black dot). Such conditions lead to stable, dry weather with clear skies, leading to warm local temperatures. The scatter plots below indicate a distinct relation between the amplitude of this pattern and the local temperature anomaly. For instance, all anomalies above 5 degrees occur on days when the amplitude of the EAF is larger than one standard deviation. On the other hand, not all days are exceptionally warm for EAF amplitudes larger than one standard deviation. The relationship between circulation and local temperature as simulated by the model is similar as in the observations. Data of all 17 simulations are plotted in the model figure.

**RL4.4. Evaluation of the 2051-2100 period**

**RL4.4.1 Changes in mean and standard deviation**

The color shading in Figure 20 denotes the change in the mean and variability of the streamfunction and temperature across Europe. The mean streamfunction shows an enhanced ridge over France and deepened troughs on both sides. The variability decreases over most part of the region, except for Northwestern Europe where the variability increases. The temperature variance increases over the continent and most in the Southern part, despite the decreased streamfunction variability. The mean warming is strongest in Southern Europe as well.

**RL4.4.2 Changes in warm extremes in the Netherlands**

The same EAF analysis was applied to the 2051-2100 period. The results are plotted in Figure 22. The same blocking pattern is found in the future period as the dominant EAF. So the same circulation anomaly in future gives rise to local high temperatures. But does this pattern occur more often in future?

To answer this question we calculated the frequency distribution of the amplitude time series of the dominant EAF for both the historical and future period (upper right panel in Figure 22). The historical distribution is uni-modal and slightly skewed towards positive values (solid blue line). The change in this distribution for the future period indicates a shift of the whole distribution towards positive values (solid red line). This is consistent with the observation that the mean change (Figure 20, upper right) projects positively onto this pattern. To check whether the distribution merely shifts and does not change shape, we calculated the frequency distribution again but with the mean change removed (dashed red line). The change with respect to the historical distribution is close to zero, so the change in the distribution is indeed a mere shift. To analyse the impact of a change in the pattern, we repeated the latter two distribution calculations with the EAF pattern of the historical period (upper left panel of Figure 21) and included these as the dashed blue and stippled blue lines respectively. The calculated changes in the distribution are close to the changes when using the EAF pattern of the future period which tells us that the (very small) pattern changes did not impact the distribution. But did the relation between the pattern and the local temperature change?

The scatter plots of EAF amplitude versus temperature are shown in Figure 22 (lower panels), along with an indication of the 5% quantile, median and 95% quantile (short stripes on the axis). In the lower left plot, the mean change is included, whereas in the lower right, the mean change is subtracted. Again we verify that the frequency
distribution of the EAF merely shifts a bit towards positive values. At the same time, the relation to temperature anomalies changes; for a given EAF amplitude change, in future the local temperature responds a little stronger; more so for the higher extremes. As a result, the width of temperature distribution increases in addition to the overall shift to higher temperatures.

The reason for the stronger response of the temperature is (1) a drying of the soil leading to a reduction in the evaporative cooling for the warm extremes and (2) a reduction of the cloud cover leading to an increase in solar radiation. This is illustrated by the monthly mean August time series of temperature, evaporation, soil moisture and net surface solar radiation in Figure 23. The probability of extreme hot months increases as well as extreme dry months with low evaporation and high amounts of solar radiation. An extreme example is August 2097 of member 13 as indicated by the green dot.

**RL4.4.3 Changes in warm extremes across Europe**

We assessed the changes in the warm extremes at eight other locations: France, Spain, Greece, Romania, Russia, Poland, Scandinavia and England. The results are plotted in Figure 24.

To first address the ability of the model to capture the principal circulation structures that relate to the local temperature anomalies, we compared the ERA40 EAF patterns (left panels of Figure 24) to the model simulated patterns (second panel). In the model world, similar circulation structures are connected to the local temperature anomalies. Overall the standard deviation of the EAF amplitudes is slightly higher in the model. The EAF patterns do not change much in the future period; the patterns of the historical period are not shown but the fact that in the frequency distribution plots (right panels) the blue dashed line is close to the red dashed line proves that the patterns do not change much (as explained in the discussion of Figure 22).

For the more southern locations, the frequency distribution of the EAF amplitudes tends to be negatively skewed, whereas for the more northern locations it is positively skewed. Spain, Greece and Romania see changes in the shape of the frequency distribution of the EAF amplitudes, the other locations experience a mere shift.

For Spain and Greece, the relation between the EAF amplitude and temperature is less tight than for the other locations, indicating that other factors (circulation structures) have a stronger influence on the local temperature as compared to the other locations.

**RL4.5 Conclusions**

The ECHAM5/OMI model reproduces the observed summer circulation and variability across Europe reasonably well; similar circulation structures are related to local temperature anomalies. In future, changes in these structures and their frequency distributions appear small. Local physical processes connected to the hydrological cycle are more important for the simulated future changes in the temperature distribution, except for the most southern locations analysed (Spain and Greece).
Figure 20: Validation of June-July 500 hPa streamfunction (upper 4 panels) and surface air temperature (lower 4 panels) for the period 1958-2000. Left for ERA40 reanalysis data, right for the ECHAM5/OMI simulations. First two plots correspond to the mean field, second two to the standard deviation. Units for streamfunction are 1e-6 m²/s, for temperature degrees Celsius.
Figure 21: Dominant EAF pattern for the Netherlands for ERA40 (upper left) and ECHAM5/OMI (upper right). The amplitude of the pattern corresponds to one standard deviation [1e-6 m^2/s]. Scatter plots of positive daily mean temperature anomalies versus the amplitude of the dominant EAF for ERA40 (lower left) and ECHAM5/OMI (lower right).
**Figure 22:** Dominant EAF pattern for the Netherlands for 2051-2100 (upper left). (upper right) Frequency distribution of the amplitude of the dominant EAF for the historical period (solid blue), the change in this distribution in the future period 2050-2100 (solid red), the change in this distribution in the future period but with removal of the mean change (dashed red), the change in this distribution in the future but using the EAF pattern of the historical period (dashed blue) and the change in this distribution in the future but using the EAF pattern of the historical period and with removal of the mean change (stippled blue). (lower left) Scatterplots of the temperature anomaly versus the dominant EAF amplitude for the historical period (blue) and the future period (red). (lower right) Same but with removal of the mean change. The colored stripes on the axis indicate the 5% quantile, the median and the 95% quantile, again with blue for the historical period, red the future.
Figure 23: Monthly mean time series for August in the Netherlands for surface air temperature (upper left), evaporation (upper right), soil moisture (lower left) and net surface solar radiation (lower right). Red crosses denote the 17 individual ensemble members, the black line is the ensemble mean, the blue line connects the values of ensemble member 13.
Figure 24: Dominant EAF for 1958-2000 ERA40 data (left panels) and for 2051-2100 model data (second column) and scatter plots of temperature anomalies versus EAF amplitudes for historical (blue) and future period (red) (third column) and frequency distribution of the EAF amplitudes (right panels, for explanation see figure 22).
References


**Ensembles Papers**

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