Atmospheric Dynamics is the Largest Source of Uncertainty in Future Winter European Rainfall

DAVID FEREDAY, ROBIN CHADWICK, JEFF KNIGHT, AND ADAM A. SCAIFE

Met Office Hadley Centre, Exeter, United Kingdom

(Manuscript received 26 January 2017, in final form 1 September 2017)

ABSTRACT

The IPCC Fifth Assessment Report highlighted large uncertainty in European precipitation changes in the coming century. This paper investigates the sources of intermodel differences using CMIP5 model European precipitation data. The contribution of atmospheric circulation to differences in precipitation trends is investigated by applying cluster analysis to daily mean sea level pressure (MSLP) data. The resulting classification is used to reconstruct monthly precipitation time series, thereby isolating the component of precipitation variability directly related to atmospheric circulation. Reconstructed observed precipitation and reconstructions of simulated historical and projection data are well correlated with the original precipitation series, showing that circulation variability accounts for a substantial fraction of European precipitation variability. Removing the reconstructed precipitation from the original precipitation leaves a residual component related to noncirculation effects (and any small remaining circulation effects). Intermodel spread in residual future European precipitation trends is substantially reduced compared to the spread of the original precipitation trends. Uncertainty in future atmospheric circulation accounts for more than half of the intermodel variance in twenty-first-century precipitation trends for winter months for both northern and southern Europe. Furthermore, a substantial part of this variance is related to different forced dynamical responses in different models and is therefore potentially reducible. These results highlight the importance of understanding future changes in atmospheric dynamics in achieving more robust projections of regional climate change. Finally, the possible dynamical mechanisms that may drive the future differences in regional circulation and precipitation are illustrated by examining simulated teleconnections with tropical precipitation.

1. Introduction

The large social and economic impacts of flooding events such as the extreme winters of 2013/14 (Huntingford et al. 2014) and 2015/16 (Scaife et al. 2017) in the United Kingdom underline the importance of understanding future trends in regional precipitation and their drivers. Anthropogenic signals were detected in the flooding of winter 2013/14 (Schaller et al. 2016; Knight et al. 2017) and autumn 2000 (Pall et al. 2011), showing the potential increased risk of further events in future. Climate model projections of precipitation for the twenty-first century are summarized in the IPCC AR5 report (Collins et al. 2013). On a global scale, precipitation is projected to rise

in line with increased global mean surface temperature with near certainty; when averaged over large spatial scales, wet regions are predicted to get wetter and dry regions to get drier as a result of thermodynamic effects (Held and Soden 2006), although this relationship may be modified for land (Byrne and O’Gorman 2015). However, projected precipitation changes show a large intermodel spread in many regions (Kent et al. 2015). Patterns of seasonal precipitation changes by the end of the twenty-first century tend toward wetter conditions in northern Europe and drier conditions in southern Europe, but the multimodel mean changes tend to be smaller than the internal variability (Fig. 12.22 in Collins et al. 2013). While it is possible that these future climate trends may genuinely be small, there may simply be a lack of model consensus, with the very

Denotes content that is immediately available upon publication as open access.

Corresponding author: David Fereday, david.fereday@metoffice.gov.uk

DOI: 10.1175/JCLI-D-17-0048.1

This article is licensed under a [Creative Commons Attribution 4.0 license](http://creativecommons.org/licenses/by/4.0/) (<http://creativecommons.org/licenses/by/4.0/>).

broad range of trends in different climate model projections averaging to something much weaker than in many of the individual projections, any one of which could be correct.

Previous attempts have been made to decompose the uncertainty in model projections. [Hawkins and Sutton \(2011\)](#) identified internal (unforced) variability as one of three sources of uncertainty in CMIP3 regional precipitation projections. The other sources are scenario uncertainty (uncertainty in the future radiative forcing imposed on the models) and model uncertainty (whereby different models produce different results for the same forcing). For European winter precipitation, they find that internal variability is the dominant source of uncertainty for the next two or three decades, while model uncertainty dominates for projections of the end of the twenty-first century, even dwarfing emission scenario uncertainty. Therefore, on long-term climate change time scales, more confident predictions of European precipitation changes, and their associated impacts, rely largely on our ability to improve climate models.

In Europe, regional precipitation is closely linked to regional modes of circulation variability, most notably the North Atlantic Oscillation (NAO). For the North Atlantic/European region, the atmospheric circulation is driven by many factors that are challenging to model such as troposphere–stratosphere interactions ([Kidston et al. 2015](#)), coupling to sea surface temperatures (SSTs; [Rodwell et al. 1999](#); [Maidens et al. 2013](#); [Fereday et al. 2008](#)), tropical precipitation ([Toniazzi and Scaife 2006](#)), and sea ice and snow ([Cohen et al. 2014](#)); future mean sea level pressure (MSLP) changes are less robust at regional scales than temperature changes [cf. Figs. 12.11 and 12.18 in [Collins et al. \(2013\)](#)]. Storm tracks are thought likely to move poleward ([Barnes and Polvani 2013](#)), leading to reduced MSLP at high latitudes and increased MSLP at midlatitudes. Additionally, dynamical poleward expansion of subtropical dry zones found in the CMIP5 RCP8.5 models ([Scheff and Frierson 2012](#)) may explain the tendency for drying in southern Europe. However, the lack of consensus for European changes means that multimodel mean changes are only significant compared to internal model variability in parts of southern Europe (Fig. 12.18 in [Collins et al. 2013](#)). This suggests that much of the uncertainty in the CMIP5 projections of European precipitation could be linked to uncertainty in regional atmospheric circulation, including not only the circulation response to climate forcing but also circulation changes due to internal variability.

It is therefore important to estimate the relative size of internal variability and uncertainty in model

response to climate forcing for circulation changes. Internal variability is irreducible, whereas uncertainty in models' forced responses could be reduced through improvements to model dynamics. [Deser et al. \(2012\)](#) argue that relatively large ensembles of model simulations may be needed to accurately estimate the forced response and internal variability. Drawing conclusions about internal variability may therefore be more challenging when using the CMIP models, where ensemble members are limited. Using a large ensemble based on a single model, [Deser et al. \(2017\)](#) show that internal variability in the NAO has a significant influence on temperature and precipitation projections for the next 50 years.

Links between circulation and precipitation changes in the Mediterranean basin have been studied by [Zappa et al. \(2015\)](#), who find that CMIP5 models showing larger changes in the North Atlantic eddy-driven jet also show larger changes in the precipitation. Circulation changes are found to account for 85% of the change in the precipitation between 1976–2005 and 2071–2100. [Santos et al. \(2016\)](#) perform an analysis of weather types and precipitation for western Europe. They find a circulation-driven increasing trend in precipitation in northwest Europe and a decreasing trend in southwest Europe.

Here we present a method for reconstructing European precipitation based on circulation, so that precipitation can be partitioned into circulation-related and non-circulation-related components. This allows us to quantify the contribution of dynamical uncertainty to the spread in projections of future precipitation in Europe. Similar approaches have been applied to model projections of European temperatures and precipitation ([Cattiaux et al. 2013, 2015](#); [Deser et al. 2017](#)), historical European temperature and precipitation ([Saffioti et al. 2016](#); [O'Reilly et al. 2017](#)), and historical temperatures for North America ([Deser et al. 2016](#); [O'Reilly et al. 2017](#)) and the Northern Hemisphere ([Smoliak et al. 2015](#)). Data and methods are summarized in [section 2](#). Results are described in [section 3](#). We first validate the method using observed MSLP and precipitation data and compare this with historical simulations from the CMIP5 ensemble. We then apply the same method to MSLP and precipitation data from CMIP5 RCP8.5 projections and measure the contribution of dynamical uncertainty to the spread in precipitation trends over the twenty-first century. We also estimate the contribution of internal variability to the spread in dynamical precipitation trends. In [section 4](#), we investigate links between tropical precipitation and European circulation as an example of the possible drivers of differences in projected model European

circulation responses, while [section 5](#) presents our conclusions.

2. Data and methods

Historical MSLP fields are taken from the European and North Atlantic Daily to Multidecadal Climate Variability (EMULATE) MSLP (EMSLP) dataset. This is a gridded dataset of daily mean MSLP based on observed data that covers the North Atlantic and European region produced in the EC-funded EMULATE project ([Ansell et al. 2006](#)). The data have a 5° resolution and cover the period 1850–2003 (the long time period may provide a more complete range of possible circulation types). MSLP anomaly fields are produced by removing a seasonally varying climatology; this climatology is produced by averaging the MSLP fields for each calendar day and then smoothing through repeated application of a 1–2–1 binomial filter. The seasonal cycle is removed to reduce its influence on the cluster classification produced. Clusters are produced from the anomaly fields using *k*-means cluster analysis ([Fereday et al. 2008](#)). This approach was found to be among the better-performing classification methods for stratifying European temperature and precipitation in a comparison by [Huth et al. \(2016\)](#). The *k*-means method assigns each field to one of a set of 30 clusters described in [Neal et al. \(2016\)](#), defined on the domain 30°W – 20°E , 35° – 70°N ([Fig. 1](#)). The choice of 30 clusters used in [Neal et al. \(2016\)](#) was made as a compromise between having too few types (so that not all important synoptic situations are represented) and too many (so that neighboring types look very similar to one another). The classification produced from the cluster analysis is extended to 2005 using daily MSLP fields taken from the NCEP–NCAR reanalysis ([Kalnay et al. 1996](#)). These additional fields are assigned to the cluster whose centroid is closest, using as a measure of distance the area-weighted sum of squared differences at each grid point.

To examine links between observed atmospheric circulation and precipitation, daily area mean precipitation time series were calculated for northern Europe (48° – 75°N , 10°W – 30°E) and southern Europe (35° – 48°N , 10°W – 30°E). See [Fig. 2](#). Daily-average precipitation data for the period 1950–2005 from the ENSEMBLES (E-OBS) precipitation dataset ([Haylock et al. 2008](#)) was used.

Given a classification of daily pressure fields and a daily time series of precipitation, we produce a reconstructed monthly precipitation time series by summing daily precipitation values from analog days for each day of the month. As an example, consider reconstructing the precipitation for January 2000. Each day of the

month belongs to one of the 30 clusters in the MSLP classification—1 January 2000 is in cluster 15, for instance. We define the analogs of 1 January 2000 as all the January days in different years that also belong to cluster 15. We choose one of these analog days at random and add that day's precipitation to the reconstructed precipitation total for January 2000. We follow the same procedure for the other days of the month to obtain a reconstructed precipitation total for the month. This total depends on the random choice of analog day for each day of the month. We therefore repeat the reconstruction process 1000 times and define the median value of the set of 1000 reconstructed monthly totals as the best estimate of the monthly total precipitation for January 2000, given knowledge of only the month's daily circulation patterns. The reconstruction time series is produced by reconstructing precipitation for each month in this way. This method assumes that precipitation values from different days of the same month are interchangeable.

Because the precipitation for any given calendar month in each year is reconstructed from the same population of possible analog days, the link between circulation and precipitation is implicitly assumed to be stationary. Changes in precipitation due to (possibly nonstationary) factors other than circulation can be examined by analyzing the residual time series; this is defined for each month as the original precipitation minus the median reconstructed precipitation for that month.

The reconstruction method is therefore applied (using the same set of 30 clusters) to simulated daily mean MSLP and precipitation fields for 23 models from the CMIP5 ensemble ([Taylor et al. 2012](#)). (An alternative approach of generating a separate set of clusters for each model was also tried, but this produced similar results.) We use daily data from the historical simulations for the 1950–2005 period and RCP8.5 scenario simulations for the 2006–99 period. The 1950–2005 period represents the overlap between the E-OBS precipitation data and the CMIP5 historical simulations. The data used are for a single ensemble member of each of the models listed in [Table 1](#). Most models only have daily data for a single ensemble member available. Daily data from the CMIP5 preindustrial control simulations are also used for the 19 models where at least 250 years of data are available (most of these simulations are at least 500 years long). Monthly mean data for a slightly different set of models are used in the analysis of teleconnections in [section 4](#). As with the observed data, model MSLP anomaly fields are produced by removing the seasonal cycle. We assign the models' daily mean MSLP fields to our set of 30 clusters, regridding the data to be consistent with the clusters' 5° grid. Each anomaly field is assigned

FIG. 1. The cluster centroids used to classify MSLP in order to reconstruct northern Europe and southern Europe area mean precipitation. Colors show cluster centroids (defined using pressure anomalies from a seasonal climatology), and black contours show absolute pressure values associated with each cluster.

to the cluster with the closest centroid, as was done for reanalysis data. This process defines an MSLP classification for each model. Daily precipitation area mean time series are produced (from the model precipitation data) for the same northern and southern European regions as were used for the observed data, and reconstructed precipitation time series are calculated. As before, this analysis allows a partitioning of the precipitation trend in each model into a dynamical component due to the effects of atmospheric circulation and a residual component due to other factors.

The robustness of the results was investigated in several ways. In each case, MSLP classifications were used to reconstruct the EOBS precipitation, and the correlations between the reconstructed and observed precipitation for each month were analyzed. To ensure that our method is insensitive to the choice of the number of clusters k , a set of observed MSLP classifications was produced for different values of k between 2 and 200. As k is increased from 2, the correlations improve, but for $k > 10$ the correlations are very similar for different k and remain high even

FIG. 2. (top) Location of northern Europe (48° – 75° N, 10° W– 30° E) and southern Europe (35° – 48° N, 10° W– 30° E) regions. (bottom) The 1981–2000 climatological means of area mean monthly precipitation totals for northern Europe and southern Europe.

for $k = 200$ (not shown). Our choice of $k = 30$ thus appears reasonable.

The correlations between observed and reconstructed precipitation are also insensitive to the source of MSLP data; this was tested using a set of 30 clusters produced for the same 30° W– 20° E, 35° – 70° N region as described above but using NCEP reanalysis data for the 1950–2005 period instead of the EMSLP data (not shown). Sets of 30 MSLP clusters were also produced (using NCEP data for 1950–2005) for the same northern Europe and southern Europe regions as were used to calculate the area mean precipitation time series. Once again, this made little difference to the observed/reconstructed precipitation correlations (not shown).

A further test of our method is to correlate monthly MSLP (from NCEP reanalysis) with the reconstructed and residual precipitation time series. As expected, the MSLP shows high correlations with the reconstructed precipitation, particularly in winter; for example, the correlation patterns with reconstructed precipitation for northern and southern Europe resemble the positive and negative phases of the winter NAO, respectively (not shown). MSLP correlations with residual precipitation are much lower, suggesting that the method captures most of the dynamical component of precipitation.

Note that while the observed EOBS precipitation data are for land only, the model precipitation area means are calculated over the whole domain and therefore include both land and ocean (we do this to reduce noise in the precipitation time series, given the coarse resolution of some of the CMIP5 models). To ensure that the different land/ocean coverage is unimportant for our validation of the reconstruction method, we analyzed NCEP reanalysis 2 precipitation data (Kanamitsu et al. 2002) for the period 1979–2015. For both northern Europe and southern Europe, two area mean precipitation time series were produced, one with ocean grid points masked out and the other with no masking applied. The cluster reconstruction method was then applied to all time series, using the EMSLP-based classification extended to 2015. For both regions, the month-by-month correlations between the actual and reconstructed precipitation for the land-only or land and ocean time series are very similar (not shown). This suggests that the use of the land-only EOBS precipitation dataset is a valid test of the reconstruction method.

Previous authors (Deser et al. 2016; O'Reilly et al. 2017) use a different reconstruction method to partition fields into dynamical and residual components. Although a detailed comparison of the two methods is beyond the scope of this paper, we note that our method can also be used to reconstruct precipitation at each grid point, as is done in these papers. Tests with NCEP reanalysis data and EOBS precipitation suggest that producing our reconstruction in this way and then taking an area mean gives an almost identical time series to applying our reconstruction method directly to the area mean time series.

3. Results

a. Verification of the reconstruction method

We first verify that the reconstruction method can usefully reproduce the year-to-year variations in the observed precipitation time series. Figure 3 shows the

TABLE 1. Models from the CMIP5 ensemble used in the analysis.

Model	Institute	Data used
ACCESS1.0	CSIRO–BoM, Australia	Daily/monthly
ACCESS1.3	CSIRO–BoM, Australia	Daily/monthly
BCC_CSM1.1	Beijing Climate Center, China	Daily/monthly
BNU-ESM	College of Global Change and Earth System Science (CGCESS), China	Monthly
CanESM2	CCCma, Canada	Daily/monthly
CCSM4	NCAR, United States	Monthly
CESM1(BGC)	CESM, United States	Daily/monthly
CESM1(CAM5)	CESM, United States	Daily/monthly
CMCC-CESM	CMCC, Italy	Daily
CMCC-CM	CMCC, Italy	Daily/monthly
CMCC-CMS	CMCC, Italy	Daily/monthly
CNRM-CM5	CNRM, France	Monthly
CSIRO Mk3.6.0	CSIRO–QCCCE, Australia	Daily/monthly
FIO-ESM	First Institute of Oceanography (FIO), China	Monthly
GFDL CM3	NOAA/GFDL, United States	Daily/monthly
GFDL-ESM2G	NOAA/GFDL, United States	Monthly
GFDL-ESM2M	NOAA/GFDL, United States	Monthly
GISS-E2-H	NASA GISS, United States	Monthly
GISS-E2-R	NASA GISS, United States	Monthly
HadGEM2-AO	Met Office Hadley Centre, United Kingdom	Monthly
HadGEM2-CC	Met Office Hadley Centre, United Kingdom	Monthly
HadGEM2-ES	Met Office Hadley Centre, United Kingdom	Monthly
INM-CM4.0	Institute of Numerical Mathematics, Russia	Daily/monthly
IPSL-CM5A-LR	IPSL, France	Daily/monthly
IPSL-CM5A-MR	IPSL, France	Daily/monthly
IPSL-CM5B-LR	IPSL, France	Daily/monthly
MIROC-ESM	MIROC, Japan	Daily/monthly
MIROC-ESM-CHEM	MIROC, Japan	Daily/monthly
MIROC5	MIROC, Japan	Daily/monthly
MPI-ESM-LR	MPI, Germany	Daily/monthly
MPI-ESM-MR	MPI, Germany	Daily/monthly
MRI-CGCM3	MRI, Japan	Daily/monthly
MRI-ESM1	MRI, Japan	Daily
NorESM1-M	Norwegian Climate Centre (NCC), Norway	Daily/monthly
NorESM1-ME	Norwegian Climate Centre (NCC), Norway	Monthly

correlation between the original and the median reconstructed precipitation time series for observations and the set of historical CMIP5 simulations. The calculation is performed for each month of the year and uses data for the period 1950–2005. For both northern and southern Europe, the correlations show that a substantial proportion of the observed interannual variability in precipitation is accounted for by interannual changes in atmospheric circulation. In winter, correlations are typically 0.8–0.9 in both regions, whereas in summer correlations are 0.6–0.8 in northern Europe and 0.4–0.5 in southern Europe. This is consistent with the fact that winter climate variability is much more dynamical, since large-scale circulation anomalies are stronger in winter and consequently play more of a role in determining climate variability than in summer (e.g., Beck et al. 2016), especially in southern Europe. Kendon et al. (2014) find that convective events are

more important for future U.K. precipitation extremes in summer than in winter, also suggesting that large-scale circulation is more strongly linked to precipitation in winter than summer. The reconstructed precipitation series has slightly reduced interannual variability compared to the original series. For the EOBS data for northern Europe, the standard deviation of the reconstructed time series is about 0.8 of the actual standard deviation in winter and 0.5 in summer. For southern Europe, the corresponding figures are 0.8 in winter and 0.33 in summer.

The correlations between simulated and reconstructed precipitation in the CMIP5 historical model simulations strongly resemble those in observations, with the observed correlations lying within the inter-model spread for every month in northern Europe and all but one month in southern Europe. The correlations shown are well above the threshold required for

FIG. 3. Influence of atmospheric circulation on regional precipitation. (top) Thin blue lines show correlation between original and reconstructed precipitation for CMIP5 historical models for northern Europe area mean precipitation for 1950–2005; thick blue line shows multimodel mean correlation value. Thick red line shows the same quantity based on EOBS precipitation data for the same period. Dotted black line shows threshold for correlation significance at the 5% level. (bottom) As in (top), but for southern Europe area mean precipitation.

significance, apart from a few models in summer. This shows that the strength of the links between circulation and precipitation in the CMIP5 historical model simulations is similar to that seen in observations. As for observations, the correlation varies through the year, with higher values in winter months than in summer months, particularly for southern Europe.

The results show that a considerable portion of European precipitation variability in both models and observations is a result of circulation variability, especially in winter, and that our method can accurately reconstruct much of this precipitation variability. We next consider linear trends in observed and simulated European precipitation time series and their equivalent residual precipitation time series. The residual

precipitation time series is expected to have effects of circulation on precipitation largely removed, and so reflect the effect on precipitation due to non-circulation-related causes such as the thermodynamic effect of climate change.

For the period 1950–2005, the simulated CMIP5 trends in regional-mean precipitation show a wide spread, with no consensus on the sign of the trend in any month for either northern or southern Europe (not shown). Decadal variability is an important driver of European circulation and precipitation trends in the 1950–2005 period (e.g., Hurrell 1995), and links have been found between European precipitation and multidecadal Atlantic Ocean variability (Knight et al. 2006; Sutton and Hodson 2005; Hermanson et al. 2014). In the CMIP5 model simulations analyzed here, however, the sequence of phases of these modes is unlikely to be reproduced. Each model is likely to have its own representation of internal modes of Atlantic variability and their phasing in the 1950–2005 period (Knight et al. 2005), so a spread of model trends is not surprising.

Examining the intermodel spread in the residual trends, for which the effect of circulation on precipitation has been removed, shows a considerable reduction in the intermodel spread. Averaging over all months, the intermodel variance in the residual trends is 43% of the intermodel variance of the original trends for northern Europe and 59% for southern Europe. More than half of the spread in historical simulated northern European rainfall is therefore due to circulation effects.

b. Analysis of contributions to future European precipitation changes

Having shown that our methodology allows us to partition precipitation into circulation-related and non-circulation-related components, we now examine future European precipitation in the multimodel ensemble of CMIP5 RCP8.5 projections covering the period 2006–2100. The correlation between the monthly time series for model precipitation and cluster-reconstructed precipitation remains similar to that seen in the historical simulations for northern Europe, with the cluster reconstructed series slightly better able to reproduce the model precipitation in the winter months. For southern Europe the correlation remains similar in winter but is slightly further reduced in summer (not shown). The precipitation reconstruction method is therefore valid in the twenty-first century. A possible concern is that the daily precipitation data used for the reconstruction are not independent of the monthly precipitation series being reconstructed. For example, if there were a strong positive precipitation trend and, coincidentally, a cluster occurred preferentially in years close to 2100, this cluster

could appear wet because of the generally wetter years in which it occurred rather than for dynamical reasons. In practice, however, reconstructed monthly RCP8.5 precipitation time series are very similar regardless of whether daily data from the RCP8.5 or the (independent) preindustrial control simulations are used for the reconstruction (apart from a lower mean state for the control-based reconstruction), suggesting that this is not a significant issue.

Figure 4 shows (in blue) the simulated linear trends per century in precipitation for northern and southern Europe over the period 2006–2100 as a percentage of the 1981–2010 EOBS mean value for each month (as shown in Fig. 2). These trends show a very large spread: for example, in northern Europe in December, model trends range from a slight drying to an almost 60% increase in precipitation over the twenty-first century. Similarly, projected trends for southern Europe in March range from around -50% to $+40\%$. The range of the projections is considerably larger than the range of decadal variability seen over recent decades, as measured by the spread of decadal means of the EOBS precipitation data (decadal means for the decades 1951–60 to 2001–10 have a standard deviation of around 10%). Even this smaller historical level of decadal precipitation variability has been associated with large fluctuations in impacts from flooding (Kundzewicz et al. 2013; Petrow et al. 2009) and droughts (Sousa et al. 2011) in Europe. As a result, the very large range in projected future precipitation implies estimating regional climate change impacts and developing robust adaptation strategies on the basis of current climate models are likely to be very challenging.

Figure 4 also shows (in red) the model residual trends for each month after circulation effects are removed. Outside the summer months, the spread and hence uncertainty is consistently reduced relative to the original trends. We decompose the intermodel variance in the original precipitation trends (σ_{total}^2) into a component due to dynamical effects and a residual component:

$$\sigma_{\text{total}}^2 = \sigma_{\text{dyn}}^2 + \sigma_{\text{residual}}^2.$$

The proportion of variance in the original trends attributable to dynamical effects ($\sigma_{\text{dyn}}^2/\sigma_{\text{total}}^2$) is shown in Fig. 5 (top two panels). For the winter months (November to March), more than half of the intermodel spread is linked to circulation in both regions. For northern Europe, the proportion is about 60% in winter, while for southern Europe it is about 75%. There is also a circulation effect in the summer months, albeit somewhat smaller. This is again consistent with weaker links between circulation and precipitation in summer

FIG. 4. Northern and southern Europe area mean precipitation trends in each month for all CMIP RCP8.5 model projections to 2100. Trends are shown as a percentage of the EOBS 1981–2010 mean values for each month (as shown in bottom panel of Fig. 2). Boxes indicate the 25th and 75th percentiles of the multimodel distribution of trends, with the central line showing the median value. Whiskers show maximum and minimum values of the distribution.

than in winter, as shown by the summertime dip in correlations between observed and reconstructed summer precipitation (Fig. 3). The results for southern Europe are consistent with Zappa et al. (2015), who find a link between Mediterranean drying and a shift in the storm track in the CMIP5 RCP8.5 simulations. Our results suggest that atmospheric circulation is the single largest source of uncertainty in future European precipitation in the winter.

The component of the intermodel trend variance due to the dynamical precipitation component σ_{dyn}^2 can be further divided into two parts due to (i) internal variability of the atmospheric circulation in each model simulation and (ii) different dynamical forced responses in different models. We assume that the dynamical internal variability of each model is independent of its dynamical response to forcing, so that

FIG. 5. (top) The percentage of the multimodel variance in precipitation trends attributable to circulation effects (i.e., 1 minus the ratio of variance in the residual trends to the variance in the original trends, expressed as a percentage) for northern Europe. Dashed line and shaded region show median and 5th–95th-percentile range, respectively, of multimodel variance due to internal variability in dynamical component of trends. Dotted lines show 50% and 0% levels. (middle) The corresponding plot for southern Europe. (bottom) Ensemble mean original trends (solid lines) and residual trends (dashed lines) for northern and southern Europe.

$$\sigma_{\text{dyn}}^2 = \sigma_{\text{internal}}^2 + \sigma_{\text{forced}}^2.$$

Of course, only σ_{forced}^2 can be reduced through future improvements to model dynamics; it is therefore important to estimate its size relative to $\sigma_{\text{internal}}^2$. The number of available ensemble members for each CMIP5 model is inadequate to estimate the models' forced responses directly (Deser et al. 2012); however, σ_{forced}^2 can instead be inferred by using the CMIP5 preindustrial control simulations to estimate $\sigma_{\text{internal}}^2$. For each model, the precipitation is reconstructed as before to obtain the dynamical precipitation component.

For each model, we now use a moving 95-yr window to calculate a set of 95-yr linear trends (matching the 95-yr RCP8.5 duration) for the available control data (i.e., trends from years 1 to 95, 2 to 96, 3 to 97, and so on up to the end of the data). We then randomly select a single trend value for each model from its set of calculated linear trends. Finally, $\sigma_{\text{internal}}^2$ is estimated as the intermodel variance in the selected precipitation trends. This random selection procedure is repeated 10 000 times to give a distribution of estimates of $\sigma_{\text{internal}}^2$.

The method assumes that the internal variability in the control simulations is the same as in the RCP8.5 simulations. We cannot exclude the possibility that the internal variability changes in response to climate forcing. Palmer (1999) argued that climate forcing projects onto modes of natural variability, so that frequencies of recurrent preferred states of the system could be altered (Corti et al. 1999; Hsu and Zwiers 2001). In addition, the future mean state of internal modes such as the NAO could shift (Gillett and Tyfe 2013).

We therefore validate the method by applying it to a 500-yr section of the preindustrial control integration and 32 RCP8.5 members taken from the CESM Large Ensemble (Kay et al. 2015). Because each ensemble member uses the same model with the same forcing, we know that σ_{forced}^2 is zero in this case. We can therefore calculate $\sigma_{\text{internal}}^2$ directly from the spread of dynamical precipitation trends of the RCP8.5 members and compare this with the estimate derived from the control integration as described above. Figure 6 compares the 50th and 5th–95th percentile ranges of $\sigma_{\text{internal}}^2$ estimated from the control integration (red lines and shaded regions) with the “true” RCP8.5-derived values (blue lines). As can be seen, the control integration-based estimates of $\sigma_{\text{internal}}^2$ are reasonably close to the RCP8.5 values, with overestimates in some months and underestimates in others. Having demonstrated that the method gives a plausible estimate of $\sigma_{\text{internal}}^2$ in this case, we apply it to the CMIP5 models.

For the CMIP5 models, the median and 5th–95th percentile range of the estimated $\sigma_{\text{internal}}^2$ distribution is

shown by the dashed lines and shaded regions in Fig. 5 (top two panels) as a percentage of the total intermodel variance. The gaps between the solid lines and dashed lines or shaded regions therefore show estimates of σ_{forced}^2 . As can be seen, σ_{forced}^2 represents a substantial proportion of the total intermodel spread in both northern and southern Europe for many months, even for the 95th percentile estimate of $\sigma_{\text{internal}}^2$. This suggests considerable scope for reduced spread in projected precipitation trends through improvements to model dynamics.

One might expect that residual precipitation trends are linked to thermodynamic effects. Residual trends in northern Europe precipitation do indeed show positive intermodel correlations with the twenty-first century change (2071–2100 RCP8.5 mean minus 1971–2000 historical mean) in North Atlantic SST (i.e., those models with larger increases in North Atlantic SST tend to have more positive residual precipitation trends). The changes between these two 30-yr periods in the RCP8.5 scenario are typically sufficiently large that the effects of internal variability (such as the Atlantic multidecadal oscillation) are likely to be small. [This is confirmed by analyzing the twenty-first century change in North Atlantic area mean SST for 30 members of the CESM Large Ensemble. The ensemble mean twenty-first century change (2.53 K) is much greater than the ensemble standard deviation (0.057 K).] The corresponding correlations for southern Europe are less marked (not shown), perhaps because soil moisture is more important for precipitation in this region in future (Seneviratne et al. 2013).

The multimodel-mean-simulated trends in both regions show an annual cycle (bottom panel of Fig. 5). For northern Europe the average precipitation trend is for wetter conditions in winter but relatively little change in summer, while southern Europe is predicted to experience pronounced drying in summer and a relatively smaller drying in winter. Broadly speaking, the residual trends for both regions (dashed lines in the figure) show the same tendencies as the original trends. The residual trends have slightly smaller amplitudes, suggesting that circulation effects are partly responsible for some of the more extreme model precipitation projections. The principal finding in these results is that removing the circulation-related variability makes the climate projections less uncertain, rather than changing the multimodel mean projection. It is clear that the single largest component of the uncertainty in the CMIP5 precipitation projections for Europe in winter is due to uncertainty in the projection of future circulation patterns. Furthermore, a substantial part of the uncertainty is related to different

FIG. 6. Month-by-month estimates of $\sigma_{\text{internal}}^2$ derived from the CESM Large Ensemble. Blue line shows $\sigma_{\text{internal}}^2$ calculated directly from the trends in the dynamical component of precipitation from the RCP8.5 members. Red line and red shaded region show median and 5th–95th-percentile range, respectively, of $\sigma_{\text{internal}}^2$ estimated from trends in preindustrial control run dynamical precipitation.

dynamical responses to forcing in different models and can thus potentially be reduced.

4. Links to tropical precipitation changes

The above results identify the response of atmospheric circulation to climate change as an important source of uncertainty in projected European precipitation. This implies that if prediction of future precipitation is to be improved, then the dynamical reasons for the different circulation changes found in models need to be understood and errors rectified. The range of possible processes driving intermodel differences is broad. This might include, for example, interactions between midlatitude cyclones and the midlatitude ocean (O'Reilly et al. 2016) or the effects of resolution. Further, there are numerous diverse influences from remote forcing factors in both the tropics (e.g., Manola et al. 2013) and high latitudes (e.g., Cohen et al. 2014). In this section we discuss the possibility of differences in the way remote influences from the

tropics could change as an example of a potential source of intermodel spread in projections of European circulation. The aim is not to provide a detailed study of this complex area but rather to illustrate the type of analyses that might lead to improved model agreement.

At interannual time scales, part of the variability in the phase of the winter NAO has been linked to teleconnections with tropical rainfall and heating anomalies (e.g., Hoskins and Karoly 1981; Sardeshmukh and Hoskins 1988; Manola et al. 2013; Gollan et al. 2015). Tropical heating leads to upper-level divergence, causing the formation of Rossby waves preferentially in regions of strong vorticity gradients associated with the subtropical jet stream. Depending on their wavelength and location, some of these waves are able to propagate to the extratropics, where they are associated with MSLP anomalies and can influence extratropical modes such as the NAO (Scaife et al. 2016). In principle, these teleconnections could provide a link to tropical regions that are themselves projected to change in different

FIG. 7. Intermodel correlation coefficients of RCP8.5 linear trends in DJF NAO index and DJF rainfall at each grid point. Stippling denotes grid points where the correlation is statistically significant at the 5% level. Boxes indicate regions discussed in the text.

ways in the range of current climate change models. Specifically, some of the dynamical uncertainty in future climate projections of the NAO might be traced to differences in changes in tropical rainfall.

Figure 7 shows intermodel correlations between 2006–2100 linear trends in the winter NAO index (calculated as the MSLP gridbox difference between the Azores and Iceland) and precipitation. Positive correlations are seen over northern Europe, indicating that (as expected) those CMIP5 models showing a stronger trend toward the positive phase of the NAO also tend to have higher northern European rainfall. Together with the negative correlations seen over southern Europe, this confirms that the NAO has a strong influence on precipitation in its own region of influence. There are also substantial correlations in several other regions, however. We focus on correlations in three regions, highlighted in Fig. 7. The two boxes in the western tropical and subtropical Atlantic are in a region of active Rossby wave formation, and ray-tracing analysis suggests that waves initiated in this region are well able to propagate into the North Atlantic and Europe and potentially influence the NAO (Scaife et al. 2016). The third region examined is the central Indian Ocean. Precipitation change in this region has also been suggested as a driver of NAO variability (e.g., Hoerling et al. 2001, 2004; Selten et al. 2004). In each region, the correlation between the trend in area mean precipitation and the NAO trend is significant at the 5% level, with correlations of 0.53, -0.46 , and 0.56 for the subtropical Atlantic, tropical Atlantic, and central Indian Oceans, respectively.

The intermodel differences in projected twenty-first-century NAO change therefore show significant correlation with tropical rainfall in regions identified as important for driving NAO variability on seasonal to decadal time scales. This suggests that differences in tropical rainfall responses to increasing greenhouse gases may be a cause of differences in the projected winter NAO. Further work, likely including suites of climate model experiments, would be required to fully test this hypothesis and quantify the size of the effect. Nevertheless, this analysis signposts a potential way to understand the dynamical origin of the model spread in the circulation and hence European rainfall. It suggests a research focus on understanding and reconciling differences in projections of tropical Atlantic and Indian Ocean rainfall could be of direct benefit to understanding and reducing uncertainty in European precipitation change.

5. Conclusions

We have developed a technique to reconstruct precipitation from atmospheric circulation data to gauge how much of the substantial uncertainty in future model projections of European precipitation is a result of uncertainty in projected circulation patterns. Classification of daily MSLP fields derived using a clustering technique allows us to reconstruct average monthly precipitation for northern and southern Europe. The reconstructed precipitation series are strongly correlated with the original precipitation, showing that much of the precipitation variability is related to large-scale

circulation patterns, particularly in winter. Residual precipitation, obtained by removing the reconstructed precipitation series from the original precipitation, is computed for climate change projections from a range of models and shows substantially less intermodel spread in twenty-first-century trends than the raw data. In the winter months, most of the uncertainty in projected precipitation trends is thereby attributable to atmospheric circulation, with sizable uncertainty reductions at other times of the year. Furthermore, estimates of the proportion of this dynamical uncertainty due to internal variability are generally fairly small. We conclude that a substantial part of the uncertainty in model projections of twenty-first-century European precipitation is due to model uncertainty in the dynamical response to climate forcing, and therefore potentially reducible. Our results are somewhat different from those of Deser et al. (2017), who find winter NAO internal variability to be an important influence on surface climate for the next 50 years. However, comparing the internal variability for each model (as measured by the variance of the control simulation dynamical precipitation trends) we find that the CESM model used by Deser et al. (2017) is at the upper end of the CMIP5 models, particularly for northern Europe (not shown). This would help to explain the differences in our results.

Our findings imply that greater focus needs to be placed on understanding the dynamical responses of the atmosphere to climate change alongside more traditional approaches, such as attempts to constrain global climate sensitivity. Currently, the range of climate model projections of European precipitation is very broad. The width of this range makes it harder to produce estimates of future climate change impacts and hence climate change adaptation strategies that are as robust as society demands (Murphy et al. 2009). It is therefore of high importance to reduce these uncertainties. Our results suggest that differences in model circulation responses to climate forcing account for a substantial proportion of the uncertainty. Furthermore, the CMIP5 models may not correctly estimate the relative contributions of internal variability and forced response. Eade et al. (2014) find underconfidence in multiyear forecasts of North Atlantic region pressure, so that ensemble members overestimate the fractional contribution of unpredictable variability. If this also applies to the CMIP5 simulations, the possible reduction of the uncertainty in projected future rainfall through improvements to model dynamics would be even larger than suggested by our analysis.

Understanding the origins of the differences in models' dynamical response to forcing is therefore

crucial. Despite the dynamical complexity of the problem, there are potentially ways in which it could be tackled. We illustrate this by showing that the relationship between intermodel differences in the winter NAO and differences in tropical rainfall resembles links between these regions identified in shorter-term climate variability. This suggests that part of the difference between models' projected NAO changes is related to differences in their projected tropical rainfall. More detailed studies of this kind could elucidate the causes of the spread in regional circulation change and point to model improvements that result in more robust projections. Further encouragement comes from recent improvements in models' ability to predict modes of regional circulation, such as the NAO, on seasonal time scales (Scaife et al. 2014). This suggests that some models can at least partly represent the principal climate dynamical influences on the extratropical atmospheric circulation.

Acknowledgments. We thank John Mitchell and Stephen Belcher for their helpful comments and suggestions. Comments from three anonymous reviewers helped improve the paper. This work was supported by the Joint U.K. BEIS/Defra Met Office Hadley Centre Climate Programme (GA01101).

REFERENCES

- Ansell, T., and Coauthors, 2006: Daily mean sea level pressure reconstructions for the European–North Atlantic region for the period 1850–2003. *J. Climate*, **19**, 2717–2742, <https://doi.org/10.1175/JCLI3775.1>.
- Barnes, E., and L. Polvani, 2013: Response of the midlatitude jets, and of their variability, to increased greenhouse gases in the CMIP5 models. *J. Climate*, **26**, 7117–7135, <https://doi.org/10.1175/JCLI-D-12-00536.1>.
- Beck, C., A. Philipp, and F. Streicher, 2016: The effect of domain size on the relationship between circulation type classifications and surface climate. *Int. J. Climatol.*, **36**, 2692–2709, <https://doi.org/10.1002/joc.3688>.
- Byrne, M., and P. O'Gorman, 2015: The response of precipitation minus evapotranspiration to climate warming: Why the wet-get-wetter, dry-get-drier scaling does not hold over land. *J. Climate*, **28**, 8078–8092, <https://doi.org/10.1175/JCLI-D-15-0369.1>.
- Cattiaux, J., H. Douville, and Y. Peings, 2013: European temperatures in CMIP5: Origins of present-day biases and future uncertainties. *Climate Dyn.*, **41**, 2889–2907, <https://doi.org/10.1007/s00382-013-1731-y>.
- , —, R. Schoetter, S. Parey, and P. Yiou, 2015: Projected increase in diurnal and interdiurnal variations of European summer temperatures. *Geophys. Res. Lett.*, **42**, 899–907, <https://doi.org/10.1002/2014GL062531>.
- Cohen, J., and Coauthors, 2014: Recent Arctic amplification and extreme mid-latitude weather. *Nat. Geosci.*, **7**, 627–637, <https://doi.org/10.1038/ngeo2234>.
- Collins, M., and Coauthors, 2013: Long-term climate change: Projections, commitments and irreversibility. *Climate Change*

- 2.013: *The Physical Science Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 273–309.
- Corti, S., F. Molteni, and T. Palmer, 1999: Signature of recent climate change in frequencies of natural atmospheric circulation regimes. *Nature*, **398**, 799–802, <https://doi.org/10.1038/19745>.
- Deser, C., A. Phillips, V. Bourdette, and H. Teng, 2012: Uncertainty in climate change projections: The role of internal variability. *Climate Dyn.*, **38**, 527–546, <https://doi.org/10.1007/s00382-010-0977-x>.
- , L. Terray, and A. Phillips, 2016: Forced and internal components of winter air temperature trends over North America during the past 50 years: Mechanisms and implications. *J. Climate*, **29**, 2237–2258, <https://doi.org/10.1175/JCLI-D-15-0304.1>.
- , J. Hurrell, and A. Phillips, 2017: The role of the North Atlantic Oscillation in European climate projections. *Climate Dyn.*, **49**, 3141–3157, <https://doi.org/10.1007/s00382-016-3502-z>.
- Eade, R., D. Smith, A. Scaife, E. Wallace, N. Dunstone, L. Hermanson, and N. Robinson, 2014: Do seasonal-to-decadal climate predictions underestimate the predictability of the real world? *Geophys. Res. Lett.*, **41**, 5620–5628, <https://doi.org/10.1002/2014GL061146>.
- Fereday, D., J. Knight, A. Scaife, C. Folland, and A. Philipp, 2008: Cluster analysis of North Atlantic–European circulation types and links with tropical Pacific sea surface temperatures. *J. Climate*, **21**, 3687–3703, <https://doi.org/10.1175/2007JCLI1875.1>.
- Gillett, N., and J. Fyfe, 2013: Annular mode changes in the CMIP5 simulations. *Geophys. Res. Lett.*, **40**, 1189–1193, <https://doi.org/10.1002/grl.50249>.
- Gollan, G., R. Greatbatch, and T. Jung, 2015: Origin of variability in Northern Hemisphere winter blocking on interannual to decadal timescales. *Geophys. Res. Lett.*, **42**, 10 037–10 046, <https://doi.org/10.1002/2015GL066572>.
- Hawkins, E., and R. Sutton, 2011: The potential to narrow uncertainty in projections of regional precipitation change. *Climate Dyn.*, **37**, 407–418, <https://doi.org/10.1007/s00382-010-0810-6>.
- Haylock, M., N. Hofstra, A. Klein Tank, E. Klok, P. Jones, and M. New, 2008: A European daily high-resolution gridded data set of surface temperature and precipitation. *J. Geophys. Res.*, **113**, D20119, <https://doi.org/10.1029/2008JD010201>.
- Held, I., and B. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686–5699, <https://doi.org/10.1175/JCLI3990.1>.
- Hermanson, L., R. Eade, N. Robinson, N. Dunstone, M. Andrews, J. Knight, A. Scaife, and D. Smith, 2014: Forecast cooling of the Atlantic subpolar gyre and associated impacts. *Geophys. Res. Lett.*, **41**, 5167–5174, <https://doi.org/10.1002/2014GL060420>.
- Hoerling, M. P., J. Hurrell, and T. Xu, 2001: Tropical origins for recent North Atlantic climate change. *Science*, **292**, 90–92, <https://doi.org/10.1126/science.1058582>.
- , —, —, G. Bates, and A. Phillips, 2004: Twentieth century North Atlantic climate change. Part II: Understanding the effect of Indian Ocean warming. *Climate Dyn.*, **23**, 391–405, <https://doi.org/10.1007/s00382-004-0433-x>.
- Hoskins, B., and D. Karoly, 1981: The steady linear response of a spherical atmosphere to thermal and orographic forcing. *J. Atmos. Sci.*, **38**, 1179–1196, [https://doi.org/10.1175/1520-0469\(1981\)038<1179:TSLROA>2.0.CO;2](https://doi.org/10.1175/1520-0469(1981)038<1179:TSLROA>2.0.CO;2).
- Hsu, C., and F. Zwiers, 2001: Climate change in recurrent regimes and modes of Northern Hemisphere atmospheric variability. *J. Geophys. Res.*, **106**, 20 145–20 159, <https://doi.org/10.1029/2001JD900229>.
- Huntingford, C., and Coauthors, 2014: Potential influences on the United Kingdom's floods of winter 2013/14. *Nat. Climate Change*, **4**, 769–777, <https://doi.org/10.1038/nclimate2314>.
- Hurrell, J., 1995: Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science*, **269**, 676–679, <https://doi.org/10.1126/science.269.5224.676>.
- Huth, R., C. Beck, and M. Kuerov, 2016: Synoptic-climatological evaluation of the classifications of atmospheric circulation patterns over Europe. *Int. J. Climatol.*, **36**, 2710–2726, <https://doi.org/10.1002/joc.4546>.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Kanamitsu, M., W. Ebisuzaki, J. Woollen, S.-K. Yang, and J. Hnilo, 2002: NCEP–DOE AMIP-II Reanalysis (R-2). *Bull. Amer. Meteor. Soc.*, **83**, 1631–1643, <https://doi.org/10.1175/BAMS-83-11-1631>.
- Kay, J., and Coauthors, 2015: The Community Earth System Model (CESM) Large Ensemble Project: A community resource for studying climate change in the presence of internal climate variability. *Bull. Amer. Meteor. Soc.*, **96**, 1333–1349, <https://doi.org/10.1175/BAMS-D-13-00255.1>.
- Kendon, E., N. Roberts, H. Fowler, M. Roberts, S. Chan, and C. Fowler, 2014: Heavier summer downpours with climate change revealed by weather forecast resolution model. *Nat. Climate Change*, **4**, 570–576, <https://doi.org/10.1038/nclimate2258>.
- Kent, C., R. Chadwick, and D. Rowell, 2015: Understanding uncertainties in future projections of seasonal tropical precipitation. *J. Climate*, **28**, 4390–4413, <https://doi.org/10.1175/JCLI-D-14-00613.1>.
- Kidston, J., A. Scaife, S. Hardiman, D. Mitchell, N. Butchart, M. Baldwin, and L. Gray, 2015: Stratospheric influence on tropospheric jet streams, storm tracks and surface weather. *Nat. Geosci.*, **8**, 433–440, <https://doi.org/10.1038/ngeo2424>.
- Knight, J., R. Allan, C. Folland, M. Vellinga, and M. Mann, 2005: A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophys. Res. Lett.*, **32**, L20708, <https://doi.org/10.1029/2005GL024233>.
- , C. Folland, and A. Scaife, 2006: Climate impacts of the Atlantic multidecadal oscillation. *Geophys. Res. Lett.*, **33**, L17706, <https://doi.org/10.1029/2006GL026242>.
- , and Coauthors, 2017: Global meteorological influences on the record UK rainfall of winter 2013–14. *Environ. Res. Lett.*, **12**, 074001, <https://doi.org/10.1088/1748-9326/aa693c>.
- Kundzewicz, Z., I. Pińskwara, and G. Brakenridge, 2013: Large floods in Europe, 1985–2009. *Hydrol. Sci. J.*, **58**, 1–7, <https://doi.org/10.1080/02626667.2012.745082>.
- Maidens, A., A. Arribas, A. Scaife, C. MacLachlan, D. Peterson, and J. Knight, 2013: The influence of surface forcings on prediction of the North Atlantic Oscillation regime of winter 2010/11. *Mon. Wea. Rev.*, **141**, 3801–3813, <https://doi.org/10.1175/MWR-D-13-00033.1>.
- Manola, I., R. J. Haarsma, and W. Hazeleger, 2013: Drivers of North Atlantic Oscillation events. *Tellus*, **65A**, 19741, <https://doi.org/10.3402/tellusa.v65i0.19741>.
- Murphy, J., and Coauthors, 2009: UK climate projections science report: Climate change projections. Met Office Hadley Centre Tech. Rep., 193 pp.
- Neal, R., D. Fereday, R. Crocker, and R. Comer, 2016: A flexible approach to defining weather patterns and their application in weather forecasting over Europe. *Meteor. Appl.*, **23**, 389–400, <https://doi.org/10.1002/met.1563>.
- O'Reilly, C. H., S. Minobe, and A. Kuwano-Yoshida, 2016: The influence of the Gulf Stream on wintertime European blocking. *Climate Dyn.*, **47**, 1545–1567, <https://doi.org/10.1007/s00382-015-2919-0>.

- , T. Woollings, and L. Zanna, 2017: The dynamical influence of the Atlantic multidecadal oscillation on continental climate. *J. Climate*, **30**, 7213–7230, <https://doi.org/10.1175/JCLI-D-16-0345.1>.
- Pall, P., T. Aina, D. Stone, P. Stott, T. Nozawa, A. Hilberts, D. Lohmann, and M. Allen, 2011: Anthropogenic greenhouse gas contribution to flood risk in England and Wales in autumn 2000. *Nature*, **470**, 382–385, <https://doi.org/10.1038/nature09762>.
- Palmer, T., 1999: A nonlinear dynamical perspective on climate prediction. *J. Climate*, **12**, 575–591, [https://doi.org/10.1175/1520-0442\(1999\)012<0575:ANDPOC>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<0575:ANDPOC>2.0.CO;2).
- Petrov, T., J. Zimmer, and B. Merz, 2009: Changes in the flood hazard in Germany through changing frequency and persistence of circulation patterns. *Nat. Hazards Earth Syst. Sci.*, **9**, 1409–1423, <https://doi.org/10.5194/nhess-9-1409-2009>.
- Rodwell, M., D. Rowell, and C. Folland, 1999: Oceanic forcing of the wintertime North Atlantic Oscillation and European climate. *Nature*, **398**, 320–323, <https://doi.org/10.1038/18648>.
- Saffioti, C., E. Fischer, S. Scherrer, and R. Knutti, 2016: Reconciling observed and modeled temperature and precipitation trends over Europe by adjusting for circulation variability. *Geophys. Res. Lett.*, **43**, 8189–8198, <https://doi.org/10.1002/2016GL069802>.
- Santos, J., M. Belo-Pereira, H. Fraga, and J. Pinto, 2016: Understanding climate change projections for precipitation over Western Europe with a weather typing approach. *J. Geophys. Res. Atmos.*, **120**, 1170–1189, <https://doi.org/10.1002/2015JD024399>.
- Sardeshmukh, P., and B. Hoskins, 1988: The generation of global rotational flow by steady idealized tropical divergence. *J. Atmos. Sci.*, **45**, 1228–1251, [https://doi.org/10.1175/1520-0469\(1988\)045<1228:TGOGRF>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<1228:TGOGRF>2.0.CO;2).
- Scaife, A. A., and Coauthors, 2014: Skilful long range prediction of European and North American winters. *Geophys. Res. Lett.*, **41**, 2514–2519, <https://doi.org/10.1002/2014GL059637>.
- , and Coauthors, 2016: Tropical rainfall, Rossby waves and regional winter climate predictions. *Quart. J. Roy. Meteor. Soc.*, <https://doi.org/10.1002/qj.2910>.
- , and Coauthors, 2017: Predictability of European winter 2015/2016. *Atmos. Sci. Lett.*, **18**, 38–44, <https://doi.org/10.1002/asl.721>.
- Schaller, N., and Coauthors, 2016: Human influence on climate in the 2014 southern England winter floods and their impacts. *Nat. Climate Change*, **6**, 627–634, <https://doi.org/10.1038/nclimate2927>.
- Scheff, J., and D. Frierson, 2012: Robust future precipitation declines in CMIP5 largely reflect the poleward expansion of model subtropical dry zones. *Geophys. Res. Lett.*, **39**, L18704, <https://doi.org/10.1029/2012GL052910>.
- Selten, F., G. Branstator, H. Dijkstra, and M. Kliphuis, 2004: Tropical origin for recent and future Northern Hemisphere climate change. *Geophys. Res. Lett.*, **31**, L21205, <https://doi.org/10.1029/2004GL020739>.
- Seneviratne, S., and Coauthors, 2013: Impact of soil moisture–climate feedbacks on CMIP5 projections: First results from the GLACE-CMIP5 experiment. *Geophys. Res. Lett.*, **40**, 5212–5217, <https://doi.org/10.1002/grl.50956>.
- Smoliak, B., J. Wallace, P. Lin, and Q. Fu, 2015: Dynamical adjustment of the Northern Hemisphere surface air temperature field: Methodology and application to observations. *J. Climate*, **28**, 1613–1629, <https://doi.org/10.1175/JCLI-D-14-00111.1>.
- Sousa, P., R. Trigo, P. Aizpurua, R. Nieto, L. Gimeno, and R. Garcia-Herrera, 2011: Trends and extremes of drought indices throughout the 20th century in the Mediterranean. *Nat. Hazards Earth Syst. Sci.*, **11**, 33–51, <https://doi.org/10.5194/nhess-11-33-2011>.
- Sutton, R., and D. Hodson, 2005: Atlantic Ocean forcing of North American and European summer climate. *Science*, **309**, 115–118, <https://doi.org/10.1126/science.1109496>.
- Taylor, K., R. Stouffer, and G. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, <https://doi.org/10.1175/BAMS-D-11-00094.1>.
- Tonizzo, T., and A. Scaife, 2006: The effect of non-linearity on winter ENSO teleconnections over Europe. *Geophys. Res. Lett.*, **33**, L24704, <https://doi.org/10.1029/2006GL027881>.
- Zappa, G., B. Hoskins, and T. Shepherd, 2015: The dependence of wintertime Mediterranean precipitation on the atmospheric circulation response to climate change. *Environ. Res. Lett.*, **10**, 104012, <https://doi.org/10.1088/1748-9326/10/10/104012>.