Multi-annual droughts in the English Lowlands: a review of their characteristics and climate drivers in the winter half year.

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Abstract

The English Lowlands is a relatively dry, densely populated region in the southeast of the UK in which water is used intensively. Consequently, parts of the region are water-stressed and face growing water resource pressures. The region is heavily dependent on groundwater and particularly vulnerable to long, multi-annual droughts, primarily associated with dry winters. Despite this vulnerability, the atmospheric drivers of multi-annual droughts in the region are poorly understood, an obstacle to developing appropriate drought management strategies, including monitoring and early warning systems. To advance our understanding, we assess known key climate drivers in the winter half-year (October-March), and their likely relationships with multi-annual droughts in the region. We characterise historic multi-annual drought episodes back to 1910 for the English Lowlands using various meteorological and hydrological datasets. Multi-annual droughts are identified using a gridded precipitation series for the entire region, and refined using the Standardized Precipitation Index (SPI), Standardized Streamflow Index (SSI) andStandardized Groundwater level Index (SGI) applied to regional-scale river flow and groundwater time series. We explore linkages between a range of potential climatic driving factors and precipitation, river flow and groundwater level indicators in the English Lowlands for the winter half-year. The drivers or forcings include El Niño-Southern Oscillation (ENSO), the North Atlantic Tripole Sea Surface Temperature (SST) pattern, the Quasi-Biennial Oscillation (QBO), solar and volcanic forcing and the Atlantic Multi-decadal Oscillation (AMO). As expected, no single driver convincingly explains the occurrence of any multi-annual drought in the historical record. However, we demonstrate, for the first time, an association between La Niña episodes and winter rainfall deficits in some major multi-annual drought episodes in the English Lowlands. We also show significant (albeit relatively weak) links between ENSO and drought indicators applied to river flow and groundwater levels. We also show that some of the other drivers listed above are likely to influence English Lowlands rainfall. We conclude by signposting a direction for this future research effort.

1 Introduction

From 2010 until early 2012, a protracted drought affected much of the central and southern UK. Following one of the driest two-year sequences on record (Kendon et al., 2013), the drought had become severe by March 2012; river flows and groundwater levels were lower in many areas than at the equivalent time in 1976, the benchmark drought year for the region (Rodda and Marsh, 2011) and water use restrictions were implemented across the drought-affected areas. The outlook for summer 2012 was distinctly fragile, but exceptional late spring and summer rainfall terminated the drought and prevented a further deterioration in conditions. In the event, widespread flooding developed (Parry et al., 2013).

While the impact of the drought on water resources was not as extensive as feared, due to its sudden cessation before the summer, it had major impacts on agriculture, the environment and recreation (Kendon et al. 2013; Environment Agency, 2012). The 2010-2012 drought brought into focus the vulnerability of the lowland areas of south and east England to drought. This region, hereafter referred to as the English Lowlands (Fig 1), includes the driest areas of the UK. It has a relatively low annual average rainfall: a 1961-1990 areal average of 680mm, with <600mm being common in the east of the region. The English Lowlands contains some of the most densely populated areas of the UK (including London) and, correspondingly, the highest concentrations of commercial enterprise and intensive agriculture; many parts of the region are already water-stressed (Environment Agency, 2009). The south and east of England is underlain by numerous productive aquifers (Fig 1), and is highly dependent on groundwater resources, with up to 70% of the water supply being from groundwater (Environment Agency, 2006). The region is particularly vulnerable to multi-annual droughts which are typically associated with protracted rainfall deficiencies in the winter half-year, leading to the limited recharge of aquifers. The 2010-2012 drought was similar to previous multi-annual droughts in the English Lowlands, such as those in 2004-2006 and in the 1990s (1988-1992 and 1995-1997). These also caused major water shortages, with significant ecological impacts (Marsh et al., 2007).

Whilst current water management in the English Lowlands presents many challenges, such issues are likely to become much more pressing. Water exploitation is likely to intensify, given anticipated increases in population and urban development (Environment Agency, 2009). The region is projected to become appreciably warmer and drier later this century if greenhouse gas concentrations increase as expected (e.g. Murphy et al., 2008), leading to decreased summer river flows (e.g. Prudhomme et al. 2012), decreased groundwater levels (e.g. Jackson et al., 2011) and an accompanying increase in the severity of drought episodes (Burke and Brown, 2010). Although a decrease in summer flows is likely to increase the frequency of single-year, summer droughts (comparable with UK droughts of 1984 and 2003), there is currently very limited understanding of how climate change may influence the occurrence of longer, multi-season and multi-annual droughts.

The 2010-2012 drought highlights the need for research aimed at improving our understanding of the drivers of the multi-annual droughts that have the greatest impact on the English Lowlands. Such understanding is vital for improving resilience to drought episodes, and consequently fostering improved systems of drought management and water resources management. Building resilience importantly involves both the monitoring and early warning of drought. Early warnings will depend crucially on an enhanced understanding and monitoring of the remote drivers of droughts and a much improved ability to predict their consequences. This includes a better understanding of the propagation of meteorological drought through to the impacts on the hydrological cycle.

Previous attempts to identify atmospheric drivers of drought in the UK have been based mostly on the occurrence of key UK weather types favouring drought (e.g. Fowler and Kilsby, 2002; Fleig et al., 2011) or on links with sea-surface temperatures (SSTs) (Kingston et al., 2013). These studies have highlighted the importance of catchment properties in modulating hydrological droughts, particularly the substantial lag-times between atmospheric drivers and river flow responses in groundwater dominated catchments in southeast England. A review of efforts focused on seasonal predictability of UK hydrology is provided by Easey et al. (2006). The majority of studies have focused on trying to identify summer drought or low flows given preceding predictors (e.g. winter SSTs, NAO). Nevertheless, concurrent links between the North Atlantic Oscillation (NAO) and UK rainfall, including extremes, have long been established in the main winter months December to February (e.g. in both models and observations by Scaife et al. 2008). Via such rainfall influences, links between the winter NAO and river flows (Laizé and Hannah, 2010) and groundwater levels (Holman et al., 2009) have been established. However, comparatively few studies have addressed links between drought and factors such as the El Nino/Southern Oscillation (ENSO) that force atmospheric circulation anomalies like the NAO themselves. Most of these drivers can be skilfully predicted months in advance (Folland et al., 2012). Globally, ENSO has very extensive regional effects on drought or flooding periods (e.g. Ropelewski and Halpert, 1996). However, only limited studies have been carried out on the influence of remote forcings on hydrological drought anywhere in Europe. Pioneering studies by Fraedrich (1990, 1992, 1994), however, provided good, including dynamical, evidence for an influence of ENSO on winter atmospheric circulation and temperature and precipitation anomalies. Although ENSO influences on European climate were affected by poorer data then available, at the peak of El Nino Fraedrich observed a now accepted pattern of higher pressure at mean sea level (PMSL) over Arctic regions of Europe and lower pressure over southern UK and areas to the south. In particular, Fraedrich (1990) showed an enhanced frequency of cyclonic compared to anticyclonic Grosswetter weather types over Europe during El Nino in almost all days during January and February. During the peak of a La Nina, a somewhat weaker tendency to enhanced anticyclonic Grosswetter types was found in this region. Such results were weakened a little in reality because it was not realised at the time that very strong El Ninos affect European atmospheric circulation in a substantially different way from moderate El Ninos (Toniazzo and Scaife, 2006, Ineson and Scaife, 2008) In addition, Lloyd-Hughes and Saunders (2002) established links between ENSO and the Standardized Precipitation Index (SPI) for Europe, finding that precipitation is most predictable in spring. For the UK, Wilby (1993) demonstrated a higher frequency of anticyclonic weather types in winters associated with La Niña conditions, consistent with Fraedrich's analyses. However, while such studies have demonstrated potential links between winter rainfall and predictable climate drivers such as ENSO, no studies have established the additional link to multi-year hydro(geo)logical droughts.

In summary, while there has been a considerable research effort, no known studies have explored close to the full range of likely climate drivers on winter half-year rainfall in the English Lowlands, nor examined how these drivers manifest themselves in multi-annual meteorological droughts and propagate through to hydrological and hydrogeological systems. Given these knowledge gaps, key objectives of this study are to:

- Identify major multi-annual droughts in the English Lowlands since 1910.
- Characterise the expression of these droughts in precipitation, river flow and groundwater levels using standardised indices, and quantify the relative timing and impact of the multi-annual droughts between the different components of the terrestrial water-cycle.
- Assess a range of likely drivers of atmospheric circulation that may contribute in the winter half-year to multi-annual droughts in the English Lowlands.
- Conduct a preliminary examination of the links between these drivers and drought indicators to search for causal connections and point the way to future studies.

2. Identifying multi-annual droughts in the English Lowlands

Many studies have assessed the character and duration of historical meteorological and hydrological droughts in the UK. Strong regional contrasts in drought occurrence across the UK have been noted, with a particular contrast between upland northern and western UK, which is susceptible to short-term (6 month) summer half-year droughts, and the lowlands of south eastern UK that are susceptible to longer-term (18-month or greater) droughts (Jones et al. 1998; Parry et al. 2011). These findings reflect both the climatological rainfall gradient across the UK (see Section 2.2) and the predominance of groundwater dominated catchments in the south-east.

In an assessment of the major droughts affecting England and Wales since the early 1800s, Marsh et al. (2007) note that the most severe droughts in the English Lowlands have all been multi-seasonal events featuring at least one dry winter, substantial groundwater impacts being a key component. Partly resulting from the long duration of these events, and the inability of groundwater systems to recover between events, these authors note a tendency for multi-annual droughts to cluster, e.g. the "Long Drought" of the 1890s – 1910. Using the Self-Calibrating Palmer Drought Severity Index (PDSI), Todd et al. (2013) have recently reconstructed meteorological droughts for three sites in southeast England back to the 17th Century, and noted numerous "drought rich" and drought poor" periods. The causes of such clustering behaviour remain poorly understood, further underscoring the importance of understanding the likely climate drivers of long droughts.

Several studies have quantitatively examined historical droughts within south east UK, as part of wider classifications of droughts in the UK and beyond. Burke et al. (2010) quantified rainfall droughts in south east UK using gridded precipitation data while Parry et al. (2011) and Hannaford et al. (2011) identified major droughts in the southeast of the UK in a regionalised streamflow series. Both studies identified similar major droughts occurring in the mid-1960s, 1975-6, 1988-1992, 1995 -1997 and the early 2000s. More recently, Bloomfield and Marchant (2013) developed a groundwater drought index based on the Standardized Precipitation Index (SPI), identifying the same major droughts. However, to the authors' knowledge, no studies have focused on multi-annual droughts where rainfall, river flows and groundwater have been simultaneously studied using consistent indicators; a necessary first step in understanding the propagation of drought from meteorology to hydrology.

The following sub-sections identify multi-annual droughts in rainfall, river flows and groundwater. Severe droughts since 1910 are characterised in two ways. First (Sect 2.2), we identified major meteorological droughts in the areal average English Lowlands rainfall series using a simple approach based on long-term rainfall deficiencies. Second, we further quantify drought characteristics using standardized drought indicators (Sect 2.3). The rationale behind the separate approaches is that using the simple approach, we can identify multi-annual drought events including at least one winter period (which is not necessarily enforced with the later drought indicators), vital when considering relationships between remote drivers and English Lowlands winter rainfall. Furthermore, this approach can identify all droughts of different durations, whereas the Sect 2.3 analysis is influenced by the choice of averaging period used in the standardized indicators.

2.1 Data sets used to identify multi-annual droughts

A range of hydro-meteorological datasets have been used to identify multi-annual droughts through the historical record. For rainfall, the key dataset is a monthly 5 x 5 km resolution gridded dataset for the UK from 1910 to date, assembled using the methods of Perry and Hollis (2005a). This gridded dataset is based on interpolated rain-gauge observations taking into account factors such as topography. It forms the basis of UK rainfall statistics produced by the UK Met Office National Climate Information Centre (NCIC). We term this dataset 'NCIC Rainfall'.

The station network comprises between 200 and 500 stations covering the UK from 1910 to 1960, a step-increase to over 4000 for the 1960s and 1970s before a gradual decline to around 2500 stations by 2012. Despite the lower network density from 1910 to 1960, these data are still able to identify earlier historical droughts with considerable confidence. Long-term-average (LTA) values were obtained from a monthly 1 x 1 km resolution LTA gridded dataset for the period 1961-1990 (Perry and Hollis, 2005b).

River flow and groundwater level data were taken from the UK National River Flow Archive (NRFA) and National Groundwater Level Archive (NGLA). An NRFA regional river flow dataset for the English Lowlands is available to characterise total outflows from the region from 1961 to 2012 (Marsh & Dixon, 2012). The series is based on aggregated flows from large rivers and uses hydrological modelling to account for ungauged areas. The boundary shown in Fig. 1 was used to create the "English Lowlands" NCIC rainfall and NRFA regional river flow series used here. A regional groundwater level series was also created for the English Lowlands to directly compare with the English Lowlands rainfall and river flow series – further information on the derivation of the groundwater level series is provided in Sect 2.3.

In addition to the regional English Lowlands outflow series, the flow record of the Thames at Kingston, the longest in the NRFA, from 1883 to present, was used to provide a temporal coverage comparable with that of the NCIC rainfall. The river Thames has the largest catchment in the UK (9968 km^2 at the Kingston gauging station) and constitutes 15% of the English Lowlands study area. This series has been naturalised, i.e. the flows have been adjusted to take account of the major abstractions upstream of the gauging station. It should, be noted that the homogeneity of the low flow record is compromised by changes in hydrometric performance over time (Hannaford and Marsh, 2006), although this is not likely to be unduly influential for the present study that focuses on drought indicators rather than trends over time. The longest Chalk groundwater level record (starting 1932) from the Thames catchment, the Rockley borehole series, is also used to provide a long-term picture.

2.2 Identifying major rainfall droughts in the English Lowlands

Meteorological droughts are identified from monthly rainfall deficits, calculated as the monthly observed areal mean rainfall total minus the monthly 1961-1990 LTA. These deficits were accumulated over rolling multi-month time periods from 12 to 24 months long. All rainfall deficits over 170 mm (25% of annual average rainfall) over 12 to 24-month timescales were selected to give 15 notable droughts from 1910 to 2012 lasting at least one year and encompassing at least one winter – i.e. likely to have significant impact on groundwater resources. These droughts did not necessarily have below average rainfall in all months from October-March; in some instances rainfall may also have been low during the summer half-year (April-September). Table 1 shows that two droughts just exceeded 24 months in length using this method. Fig 2 shows an example rainfall anomaly series, that for the 2010-2012 drought, which includes a few months before and after the chosen drought period to demonstrate a typical example of how drought beginning and end dates were chosen.

Meteorological droughts across the English Lowlands since 1910 identified here include 1920-1921, 1933-1934, 1975-1976, 1990-1992 and 1995-1997, consistent with earlier studies (Marsh et al., 2007) so their identification is not very sensitive to the criteria used. Of these, the 1975-1976 drought is generally regarded as a benchmark across much of England and Wales against which all other droughts are often compared (Rodda and Marsh, 2011). During only this and the 1920-1921 drought were rainfall totals below 65% of LTA over the >12 month time-scale, including all or most of a winter half-year (Table 1). The most recent historical drought of 2010 to 2012 comfortably sits as one of the most significant prolonged droughts since 1910 (Kendon et al., 2013).

We also examined how spatially coherent on average these 20 major long droughts were over the UK. There is a well known strong rainfall gradient between the English Lowlands and northwest Britain (an order of magnitude between the wettest parts of the Scottish Highlands and driest parts of East Anglia). Because of the predominance of westerly airflows interacting with western uplands, eastern lowland areas are often in rainshadow. Accordingly, periods of very wet or very dry conditions in the lowlands often differ from those in northwestern UK. The atmospheric drivers of lowland UK droughts are therefore likely to be somewhat different to those in the northwest. To demonstrate this, Fig. 3 shows correlations between rainfall in the ten climatological rainfall districts covering the UK defined by the UK Met Office and gridded NCIC rainfall data elsewhere in UK for both winter and summer half years using the 15 long drought periods listed in Table 1. Although summer is not a focus of the paper, Fig 3 shows a considerable differences between winter and summer correlation patterns. Generally, there is a greater anticorrelation between southeast UK and northwest UK rainfall in the winter half year than in the summer half year. This implies that droughts have a greater tendency to affect the UK as a whole in the summer half year than in the winter half year. Indeed, Fig 3 suggests that northwest Scotland is unlikely to be affected by drought at the same time as southeast England in the winter half year. Rahiz and New (2013) have also recently confirmed a tendency for spatially coherent meteorological droughts in southeast of England to be distinct in time from droughts in northern and western areas of UK.

2.3 Identifying major droughts in rainfall, river flows and groundwater from a hydrological perspective

In order to examine the impact of historical meteorological droughts on river flows and groundwater, consistent indicators are required to identify such drought events. A wide range of drought indicators are available (e.g. Mishra and Singh, 2010) and there is no current consensus on a single indicator appropriate for capturing the wide range of drought impacts. The Standardized Precipitation Index (SPI, McKee et al. 1993) benefits from being normalised to allow comparisons between diverse regions and through the annual cycle. The formulation of the SPI is described in detail elsewhere; in summary it consists of a normalised index obtained by fitting a gamma or other appropriate distribution to the precipitation record, where fitting is done for each calendar month to account for seasonal differences. The monthly fitted distributions are transformed to a standard normal distribution and the estimated standardised values combined to produce the SPI time series. The index is fitted to precipitation data that are typically accumulated over 3, 6, 12 and 24 month periods. The SPI concept has been extended to river flows (e.g. Shukla and Wood, 2008) but numerous variants have been proposed and there is no consensus on the distributions that should be used for normalisation (e.g. Vicente-Serrano et al., 2012). More recently, the SPI concept has been extended to groundwater level records via a Standardized Groundwater level Index, SGI (Bloomfield and Marchant, 2013). This adopts a non-parametric normal scores transformation rather than using a defined statistical distribution.

For the present study, the SPI has been applied to the English Lowlands rainfall series, and the SGI has been applied to 11 individual groundwater level records from observation boreholes within the English Lowlands region. These are: Ashton Farm, Chilgrove House, Dalton Holme, Little Bucket Farm, Lower Barn, New Red Lion, Rockley, Stonor House, Therfield Rectory, Well House Inn and West Dean (see Bloomfield and Marchant (2013) for more information on these groundwater records). The groundwater hydrographs have been averaged to create a regional SGI series of English Lowlands groundwater levels. Unlike the SPI, the SGI is not applied to time series that have to be accumulated over a range of durations, because groundwater level and river flow exhibits autocorrelation or 'memory' which implies that a degree of accumulation is inherent in each monthly value. The same methodology was also applied to the English Lowlands regional river flow series (henceforth referred to as Standardized Streamflow Index, SSI). Whilst the SGI was developed primarily for groundwater, its formulation is also highly appropriate for river flows – particularly in the English Lowlands where a substantial proportion of the runoff comes directly from stored groundwater. As with groundwater levels, monthly river flows were not accumulated over a range of periods to produce the SSI for river flow.

Standardized Indices were calculated for English Lowlands regional river flow and regional groundwater levels, and monthly SPI was calculated for all accumulation periods from months 1 to 24 (i.e. SPI_1 to SPI_{24}). Figure 4a shows a heatmap of the correlation between lagged English Lowlands river flow (as SSI) and English Lowlands precipitation (as SPI_1 to SPI_{24}). The maximum correlation of 0.79 occurs for lag zero between the two time series and for a precipitation accumulation period of 3 months. Figure 4b is a similar heatmap of lagged English Lowlands mean groundwater levels (as SGI) and English Lowlands precipitation (as SPI_1 to SPI_{24}). The maximum correlation is 0.82, also for lag zero, but only for a longer precipitation accumulation period of 12 months. In summary, the highest correlations between SSI and SPI and between SGI and SPI are associated with concurrent time series, although correlations >0.75 between SGI and SPI are also seen at lags of a few months.

Figure 5 shows, for the English Lowlands, SPI rainfall series for several accumulation periods and the corresponding SSI and SGI river flow and groundwater series. Fig 6 shows the English Lowlands rainfall (SPI) series and equivalent series for the long Thames (SSI) record, and the Rockley borehole (SGI). Both figures demonstrate good agreement between the meteorological droughts and associated river flow and groundwater droughts – with some expected delays for the onset of given hydrological drought events, demonstrating the propagation between the meteorological and groundwater droughts in particular. Fig 6 also shows very good agreement between the severity of the major rainfall droughts identified independently in Sect. 2.2, suggesting that these long duration events indeed had an identifiable and considerable impact on river flows and groundwater in the English Lowlands. However, a cluster of hydrological drought events in the mid-1950s, not identified in Sect 2.2., is also apparent in Fig 6. The magnitude of the SPI/SGI/SSI anomalies in this period are not as great, but the duration is notable. Overall, these analyses demonstrate the strong link between meteorological droughts and their manifestation in hydro(geo)logical responses but they also demonstrate some differences between the two, as expected. From this it is inferred that the major long meteorological droughts identified in Table 1, and the various hydrological drought metrics used to characterise them, provide a good basis for establishing links between potential climate drivers and the major historical droughts experienced in the English Lowlands. Nevertheless, links between the remote drivers of meteorological and groundwater hydrological droughts in particular are not expected to be identical, and the lag times identified above should be considered in interpreting these relationships.

3. Climate drivers of meteorological drought in the English Lowlands

This section considers the evidence for potential forcing factors for multi-annual meteorological droughts in the English Lowlands. We selectively extend published results on the forcing of core winter atmospheric circulation anomalies, and rainfall where this exists, to the winter half-year (October-March). We show results for atmospheric circulation in a global context, and for rainfall most of western Europe, to provide the large-scale context that is appropriate to understanding forcings by remote drivers. By driving or forcing factor we mean a physical factor external to, or within, the climate system that tends to force atmospheric circulation and rainfall responses over the North Atlantic/European region in winter. We do not regard atmospheric circulation anomalies as forcing factors in this paper, though they are of course the immediate causes of anomalies of surface climate.

A necessary first-step in linking driving factors with rainfall anomalies is to consider their influence on PMSL. Thus English Lowlands rainfall anomalies on seasonal time scales are relatively highly linearly correlated with the simultaneous PMSL anomaly over the English Lowlands. Averaged over the six month winter half-year, PMSL anomalies are an especially good indicator of rainfall anomalies, the correlation between simultaneous PMSL anomalies and rainfall anomalies being

-0.78 over the period 1901-2 to 2011-12 (61% of explained rainfall variance), or -21 mm/hPa averaged over the English Lowlands. *For the English Lowlands in the winter half-year, the key to* *forecasting rainfall is skilfully forecasting PMSL anomalies averaged over the English Lowlands.* This is approximately the same as counting the relative number of cyclonic and anticyclonic days, indicating that winter mean English Lowlands flow vorticity could add some extra skill to PMSL alone. Jones et al. (2014) discuss controls on seasonal southeast England rainfall in such terms, although they do not use mean PMSL anomalies directly. However, in western regions of the UK, forecasting PMSL may not be enough; atmospheric circulation patterns like the NAO are likely to be important because near surface anomalous wind direction and speed quite strongly affect rainfall there (Jones et al., 2014).

Folland et al (2012) reviewed the influences of the then-known forcing factors in winter on European temperature and rainfall, mainly for December to February or March, and concluded that the climate models current at the time underestimated potential temperature and probably rainfall predictability. Forcing factors investigated included the El Niño-Southern Oscillation (ENSO), North Atlantic sea surface temperature (SST) patterns, the quasi-biennial oscillation (QBO) of equatorial stratospheric winds, major tropical volcanic eruptions and increasing greenhouse gases. Since that paper, physically-based influences of solar variability on winter climate have been discovered (e.g. Ineson et al., 2011, Scaife et al., 2013). Postulated influences of recently reducing Arctic sea ice extent on winter European atmospheric circulation remain unclear and are not discussed further (Cohen et al, 2014) but may still exist.

Recently, a much higher level of real-time forecast skill for the NAO has been demonstrated by Scaife et al. (2014a) for the core winter months of December-February for UK and Europe using Glosea 5, a version of the latest Met Office climate model, HadGEM3 (Maclachlan et al., 2014). Scaife et al. (2014a) show that this new level of skill reflects many of the factors reviewed by Folland et al. (2012), though not La Niña, and that none are dominant, confirming that a multivariate forcing factor approach is needed to understand interannual climate variations in the winter half-year. However, significant rainfall skill for UK regions was not shown. To investigate drivers of English Lowlands rainfall for the winter half-year, we use several data sets. These include the global 0.5^{o} x 0.5^{o} rainfall data of Mitchell and Jones (2005), PMSL data of Allan and Ansell (2006), 300hPa and PMSL data from the Twentieth Century Reanalysis (20CR) (Compo et al., 2011), the NCEP Reanalysis (Kalnay et al., 1996) and HadISST1 sea surface temperature data (Rayner et al., 2003). For La Niña data we use the Niño 3.4 index using a combination of the Kaplan et al. (1998) SST analysis to 1949 and the Reynolds et al. ERSSTv3b analysis from 1950 (updated from Reynolds et al., 2002), henceforth KRSST. Other driving data include annual total solar irradiance up to 1978 from Prather et al (2014), interpolated to monthly values, with measured monthly values from 1979 (Fröhlich,2006), May North Atlantic SST Tripole data (Rodwell and Folland, 2002, Folland et al., 2012), the Atlantic Multidecadal Oscillation (AMO) (Parker et al., 2007), stratospheric volcanic aerosol loadings (Vernier et al., 2011) and the QBO (Naujokat, 1986). For English Lowlands rainfall, we have created a combined NCIC and Mitchell et al (2005) time series from 1901-2012, regressing Mitchell et al data against the NCIC data set regarded as the primary set to extend the latter back to 1901.

In the following sections, we discuss atmospheric circulation and rainfall anomaly forcing in the winter half-year due to ENSO, the North Atlantic Tripole SST anomaly, the QBO, tropical volcanoes, solar effects and the AMO.

3.1 ENSO

Toniazzo and Scaife (2006) showed how El Niños (associated with significantly warmer than normal SST in the tropical east Pacific) affect winter, mainly January-March, extratropical Northern Hemisphere atmospheric circulation and temperature. The character and physical causes of the influences differ between moderate and strong El Niños (Ineson and Scaife, 2008). Moderate El Ninos appear to influence winter extratropical Northern Hemisphere climate through a stratospheric mechanism, whereas very strong El Ninos force a wave train through the troposphere from the tropics (Ineson and Scaife, 2008) giving very different patterns of winter atmospheric circulation response. Folland et al. (2012), their Fig 7b, show that the overall effect of El Niño on English Lowlands rainfall in December-February is towards modestly wetter than normal conditions, while La Niña (associated with significantly colder than normal SST in the tropical east Pacific) gives modestly drier conditions than normal conditions, consistent with the model results of Davies et al. (1997) and the observational results of Moron and Gouirand (2004). There is no evidence that strong La Niñas influence atmospheric circulation in different ways from moderate ones.

To investigate the influence of La Niña events, Fig 7a first shows the mean global SST anomaly pattern associated with La Niña events where SST averaged over the Niño 3.4 region (120^{o}W-170^{o}W, 5^{o}N-5^{o}S) has an anomaly <= -1.0^{o}C, compared to the 1961-1990 average. SST values averaging >= 1.0^{o}C above normal give a broadly opposite SST pattern. To provide dynamically consistent information about PMSL since the late 19th Century, we use median results from the 20CR. This assimilates observed PMSL and surface temperature data into a physically consistent climate model framework every 6 hours for most of the last 130 years using an ensemble of over 50 different slightly differing analyses. Fig 7b, top panel, shows mean PMSL anomalies (from 1961-1990) for La Niñas where Nino 3.4 region SST anomalies are <-0.92^{o}C for two independent epochs 1876-1950 and 1951-2009. The value -0.92C is minus one standard deviation of Nino 3.4 SSTs over 1951-2009. Both epochs show a finger of higher than normal PMSL stretching toward the southern UK, much stronger in the latter period, with lower than normal PMSL to the north. General similarities in the patterns tend to confirm the robustness of the PMSL pattern. PMSL anomalies project as expected onto the positive winter NAO in both epochs, but with higher PMSL over the south of the UK during La Niña than in the classical NAO pattern.

The central panel shows anomalies of atmospheric storminess from the NCEP Reanalysis for 1951-2013 and western European rainfall anomalies for 1901-2011. These show significantly drier than average conditions and slightly reduced storminess over the English Lowlands during La Niña. The dry anomalies over the English Lowlands average around 5 mm/month (30 mm in the winter half year) while northwest Scotland by contrast has significant slight to moderate wet anomalies exceeding 10mm/month. The average PMSL anomaly over the English Lowlands in 1951-2009 of 1.8hPa in Fig 7b corresponds to about a 38mm rainfall deficit, 11% of the 1961-1990 winter half year average of 348mm. The average effect is thus modest, as with all other individual climatic influences, though individual La Niña events can have a stronger influence. Details of the influence of La Niña on UK PMSL and rainfall vary through the winter half-year (e.g. Fereday et al, 2008), illustrated in Supplementary Information Fig S1 for each winter half-year month. Fig S1 shows no English Lowlands rainfall signal in January, though a dry signal appears to a greater or lesser extent in the remaining five months.

El Niño, by contrast, is associated with slightly wetter conditions than normal in the English Lowlands and slightly enhanced storminess (Fig 7b, bottom right). Indeed, broadly opposite PMSL anomaly and rainfall anomaly patterns can be seen in the bottom panels of Fig 7b in given locations over most of UK and Europe during moderate El Niños (0.92^{o}C< Niño 3.4 SST anomaly < 1.5^{o}C). For the relatively uncommon extreme El Niños, PMSL (Toniazzo and Scaife, 2006) and rainfall patterns change over the UK and English Lowlands (not shown).

Table 1 shows the mean winter half-year Niño 3.4 SST anomaly during each drought. No moderate to strong El Niños occurred in these droughts but there was one weak El Niño, four weak La Niñas (SST anomaly between 0.5 and 1^{o}C), seven “neutral” conditions (anomalies between +-0.5^{o}C, all here with weak negative SST anomalies) and three moderate to strong La Niñas. The mean winter half-year Niño 3.4 SST anomaly in all 15 droughts is -0.45^{o}C. Table 2 looks at the problem in another way, showing the winter half-year rainfall anomaly associated with the strongest La Niñas and noting if a Table 1 drought occurred. Many La Niñas are not associated with winter half-year components of Table 1 droughts. However the probability of a Table 1 drought occurring during the top 20 winter half-year La Niñas is nominally 0.35, compared to a chance probability of 0.15, so the probability of a severe drought is approximately doubled compared to chance. The overall English Lowlands winter half-year rainfall anomaly during all top 20 Nino 3.4 years is nevertheless weak at 25.2 mm or -0.39 standard deviations. So a doubling of the chance probability is worth noting, but La Niña is inadequate to indicate a Table 1 drought with any confidence by itself. Moreover, La Niña winters can occasionally behave very far from expectation. The clearest example is 2000-1, the wettest winter half-year in this record at 43 mm/month but accompanied by a weak La Niña with an SST anomaly of -0.70°C. This very cyclonic winter may have been caused by the overriding influence of other strong forcings, especially in October-December (Blackburn and Hoskins, 2001).

Finally Fig 7c shows cumulative distributions of English Lowlands rainfall when Nino3.4 SST anomalies <-0.5°C and Nino 3.4 SST anomalies >0.5°C but < 1.5°C were observed. The latter is an approximate lower Nino3.4 SST limit for extreme El Ninos; these extreme years tend to be more anticyclonic over the English Lowlands so on average drier than other El Nino years. Fig 7c shows drier conditions in La Nina compared to El Nino through almost all of the cumulative probability distribution of English Lowland rainfall. A clear exception is the wettest winter half year, 2000-2001. Including the three extreme El Nino years (not shown) slightly reduces the contrast between El Nino and La Nina influences.

3.2. Other potential climate drivers for English Lowlands rainfall in the winter half-year

3.2.1 North Atlantic tripole SST anomalies

Rodwell et al. (1999) and Rodwell and Folland (2002) showed that a tripole SST pattern in the North Atlantic in December-February was associated in climate models and observations with a weak if clear physical modulation of a PMSL pattern quite like the NAO. The tripole has been the most prominent SST pattern in the North Atlantic since the 1940s. Rodwell and Folland (2002) explain why the state of the SST tripole best predicts the winter NAO in the May prior to the winter being forecast. Folland et al. (2012) extended these results to show the European December-February winter rainfall pattern predicted by the May tripole. We further extend these results to the winter half-year, though the tripole index is currently only available for 1949–2008. Despite the short data set, composite PMSL analyses for tripole indices of <-1 SD and >1 SD give widely significant results. The positive index is associated (over this period) with a positive NAO displaced slightly southwards, and the negative index with a negative NAO (Fig 8a, c), results fairly like those for December-February. Accordingly, positive values of the tripole index in May are associated with wet conditions in western UK in the following winter half-year, though only marginally wet conditions in the English Lowlands. Negative indices give a tendency to dry conditions in western UK and to some extent the English Lowlands (Fig 8b, d). In conclusion, a negative North Atlantic SST tripole index in May tends to weakly favour dry conditions in the English Lowlands in the following winter half-year.

3.2.2 Quasi-biennial oscillation of stratospheric winds

Marshall and Scaife (2009) discuss differences in atmospheric circulation and surface temperature in the extratropical Northern Hemisphere between winters (December-February) with strong lower stratospheric westerly winds near the equator at 30hPa and those with easterly winds at that level. These winds vary with a period of between two and three years and are known as the quasi-biennial oscillation (QBO). The easterly QBO tends to increase North Atlantic blocking, with a negative NAO, in December-February while the westerly QBO mode is associated with a positive NAO. Mechanisms by which equatorial stratospheric QBO winds influence the lower winter extratropical troposphere are partly understood; Folland et al. (2012) give references. Folland et al. (2012) show precipitation anomalies for +1SD of the QBO signal but these are weak over the UK and Europe. The QBO can now be reliably forecast a year or more ahead (Scaife et al., 2014b).

Fig 9 illustrates global PMSL and rainfall anomalies over UK and nearby Europe associated with strong easterly and westerly QBO winds at 30hPa in the winter half-year. Because strong easterly QBO winds are substantially stronger than strong westerly QBO winds, we compare PMSL and rainfall for the most easterly 15% of all winter half-year QBO winds (top panels) and the most westerly 15% (bottom panels). A value of 15% is selected because although the influence on atmospheric circulation of the most westerly 10% and 10%-20% of QBO winds is similar, the easterly influence weakens below 15%. Strong easterly QBO conditions are indeed associated with blocked conditions in the winter half-year and strong westerly conditions with a positive NAO as for December-February. However PMSL is near normal for westerly QBO conditions over the English Lowlands giving no rainfall signal (bottom right). Strong easterly QBO winds tend to give a small negative PMSL anomaly over the English Lowlands with modestly wetter than average conditions (bottom left panel). So the QBO appears to have only a small influence on English Lowlands winter half-year mean rainfall. However, Fig 9 shows that strong easterly or westerly phases of the QBO quite strongly and symmetrically affect winter atmospheric circulation over the North Atlantic. Interacting with other forcing factors, QBO influences might have more importance for English Lowlands winter rainfall than this analysis suggests.

3.2.3 Major tropical volcanic eruptions

The winter (December-February) rainfall patterns associated with major tropical volcanic eruptions were shown by Folland et al (2012). Major tropical volcanic eruptions are uncommon and tend to force the positive westerly phase of the NAO in winter (e.g. Robock, 2000, Marshall et al., 2009). Wetter than normal conditions are seen in northern Scotland with slightly drier than normal conditions further south and over the English Lowlands (Fig 5 of Folland et al., 2012). Further analysis is beyond the scope of this paper. Although climate models often have difficulty with this relationship, the main cause of the increased westerly phase of the NAO is thought to be an increase in the temperature gradient in the lower stratosphere between the tropics and the Arctic. This is caused by warming of the lower stratosphere by absorption of upward long wave radiation from the troposphere and surface by the volcanic aerosols (mainly tiny sulphuric acid particles) where heating is much greater in the tropics (Robock, 2000). The resulting increased temperature gradient between the tropics and the polar regions favours stronger extratropical westerly winds in the lower stratosphere through the change in the geostrophic balance. In turn enhanced extratropical tropospheric westerly winds result through wave-mean flow interaction, a dynamical mechanism only partly understood (e.g. Perlwitz and Graf, 1995).

3.2.4 Solar effects

Solar effects on North Atlantic climate have identified in observations for winter (December-February) for Europe (e.g. Lockwood et al., 2010). Ineson et al. (2011) carried out model experiments with a vertically highly resolved model extending to the lower mesosphere to show that ultraviolet solar radiation variations associated with the 11 year solar cycle of total solar irradiance (TSI) modulate the Arctic Oscillation and NAO and thus winter blocking over UK through stratospheric-tropospheric interactions. Thus stronger solar ultraviolet radiation near the maximum of the solar cycle favours the westerly positive phase of the NAO over UK and weaker radiation at solar minimum favours blocking, easterly winds and the negative phase of NAO. Ineson et al (2011) showed that the mechanism for these effects starts in the lower mesosphere or stratosphere. Here, for example, reduced ultraviolet radiation at solar minimum causes a decrease in ozone heating. This cooling signal peaks in the tropics; so opposite to the volcanic forcing influence described above, this decreases the tropics to polar region stratospheric temperature gradient. This leads to weaker stratospheric winds as the geostrophic balance changes. These reduced winds propagate downward into the troposphere through wave-mean flow interaction to give a more negative or easterly phase than average NAO. Scaife et al. (2013) also showed that solar modulation of the NAO feeds back onto the North Atlantic SST Tripole. This in turn influences the winter atmospheric circulation which feeds back onto the SST tripole etc. As a result,te a maximum westerly positive NAO winter atmospheric circulation response occurs 1-4 years after solar maximum and a maximum easterly negative phase of the NAO occurs 1-4 years after solar minimum.

We have carried out a preliminary study for the longer October-March period. Mean PMSL anomalies in the Atlantic sector tend to be fairly consistent at or near solar maximum, but less consistent and weak around solar minimum. So we confine our results to high values of TSI. Fig 10 shows global PMSL and UK and European rainfall anomalies for winter half-year lagged by one year on average compared to the highest 20% of values of TSI over 1948-2011. A modest, significant, cyclonic anomaly occurs west of the UK with a significant if small tendency to wetter than normal conditions in the English Lowlands. The highest 25% of TSI values gives much the same result. Some studies suggest that the QBO and solar cycle phases may interact to influence North Atlantic winter atmospheric circulation (Anstey and Shepherd, 2014) in a more complex way, so this could be a topic for the future.

3.2.5 The Atlantic Multidecadal Oscillation

The AMO is likely to be both a natural internal variation of the North Atlantic Ocean (Knight et al, 2005) and anthropogenically forced (Booth et al., 2012). In a model study, Knight et al. (2006) showed influences of the model AMO on UK seasonal climate, indicating a marked variation in the effects of the AMO between three month seasons, as more recently shown by Sutton and Dong (2012) from observations. The version of the observed AMO we use here is that due to Parker et al. (2007) which reflects an associated quasi- global interhemispheric SST pattern concentrated in the North Atlantic, much as seen by Knight et al. (2005) in the HadCM3 coupled model. Fig 11 shows global

PMSL and UK and European rainfall anomalies over the common data availability period 1901-2011 for winter half-year AMO values >1 and <1 standard deviation calculated over this period. These correspond to warm and cold North Atlantic states corrected for trends in global mean sea surface temperature.. (The state in 2014 was relatively warm). The AMO varies mostly interdecadally so any AMO related climate signal is likely also mostly interdecadal. There is a significant, clear and symmetric PMSL signal over the North Atlantic region. A negative NAO is seen when the AMO is in its positive phase and a positive NAO when the AMO is negative. AMO effects on rainfall over much of UK are clearest for the negative AMO phase which favours mostly drier than average conditions in the west. Unfortunately, neither phase of the AMO provides a rainfall signal for the English Lowlands. However, Fig 11 may hide considerable variability within the winter half-year as Sutton and Dong (2012) show large differences in European climate signals between different calendar three month periods. Intraseasonal influences of the AMO on atmospheric circulation within the winter half-year require investigation.

3.3 Links between large-scale drivers and drought indicators

In this section, we explore relationships between the various potential large-scale drivers identified in Sect 3.2 and the hydrological drought indicators discussed in Section 2.

Figure 12 comprises boxplots of the various response variables for the winter half year rainfall and river flow, as well as the drought indicators (SPI, SSI and SGI) for low (<-0.5 SD) and high (>0.5 SD) values of the predictors. This figure is intended to provide an overview of possible linkages between drought relevant hydro-climatic time series and the various climate drivers discussed in this study. The driving data include Niño 3.4, the May SST tripole, the QBO, stratospheric volcanic aerosol loadings, TSI, and the AMO.

The data for the drivers and response variables in Figure 12 are mostly averaged over October-March, so that the analysis is for concurrent data. However, the groundwater SGI is averaged with a lag of two months, and is thus shown for December-May, to reflect the temporal delay in groundwater formation. Because the SPI describes rainfall accumulated over a number of preceding months, these have also been lagged compared with the drivers so as to be centred on the target period October-March. Accordingly, the SPI3 is shifted forward by 1 month, and averaged for November-April; thus the first three-month accumulation starts in September and the last ends in April. Corresponding shifts for the SPI6 and SPI12 are three and six months respectively. The TSI precedes the hydrological response variable by two years to be consistent with the findings by Scaife et al. (2013) as discussed in Sect 3.2.4. Significance levels are calculated using one-sided Welch two-sample t-tests.

As perhaps expected, given the relationships discussed in Sect 3.2, the majority of univariate relationships shown in Fig. 12 are very weak and non-significant, and the majority of individual drivers have little discernible impact on the means of the response variable. The only significant relationship for English Lowlands rainfall is with the Niño 3.4 SST anomaly. Nevertheless, there is a clear tendency for El Niños (weak, moderate and strong) to be associated with wet conditions, and higher river flows and groundwater levels, and La Niña with dry conditions and lower flows and levels, consistent with Sect 3.2 and Folland et al. (2012). As mentioned in section 3.2, a strong note of caution, and a cause of the poor significance, is that the wettest winter half year in Fig 8c, 2000-2001, is associated with a weak La Niña and not an El Niño. SPI3 shows a significant relationship with the SST tripole, which is only very weakly supported by the other variables. However, the spatial analysis shown in Fig 8 (bottom panels) suggests a stronger relationship exists for the upland north-west of the UK rather than the lowland south-east. Svensson and Prudhomme (2005) noted a positive concurrent winter (Dec-Feb) correlation between SSTs in the area corresponding to the centre of the SST tripole and river flows in northwest Britain (r=0.36), consistent with Fig. 8b and d. For river flows in southeast Britain, encompassing the English Lowlands, they found a positive concurrent winter correlation with SSTs slightly further to the south (r=0.43), partly overlapping the southernmost centre of the SST tripole.

For the majority of other potential climate drivers, the distributions of the drought indicators are typically not significantly different from one another for values >0.5 or <-0.5 SD of the respective drivers. The key finding is that no single driver is close to compellingly explaining English Lowlands rainfall, river flows or groundwater levels. Combinations of drivers are of course difficult to test with the limited observational data available.

4. Discussion

4.1 General considerations

The predictability of winter droughts in the English Lowlands is a multiple forcing problem made more difficult by the relatively small scale of the English Lowlands compared to that of atmospheric anomalies. Temperature is a small additional factor in the winter half-year for drought but much more important in summer, when high rates of evapotranspiration can exacerbate hydrological droughtIn winter, temperature could be influential in increasing the likelihood of snowfall as opposed to rainfall, which could confound links between the atmospheric drivers we have identified and precipitation, river flow and groundwater deficits. While water storage in snow/ice during the cold season can be a major influence on hydrological drought in parts of Europe (e.g. van Loon et al. 2014), generally, snowfall is limited in the English Lowlands. Some winter drought periods (e.g. 1962/63, 2010/2011) were associated with major snowfall and persistent snow cover, but typically snow makes up a modest proportion of precipitation and is a minor runoff generation component (even in cold winters) at the monthly to seasonal scale.

Our work has focused on the winter half-year, but we acknowledge that a complete discussion of the multiannual drought problem requires an investigation of the influences of remote drivers on summer half-year precipitation and temperature. Our current understanding of the drivers of atmospheric circulation in December-February over the UK and Europe has clearly improved, reflected in the new level of skill in dynamical forecasts of atmospheric circulation near UK shown by Scaife et al. (2014) mentioned in Section 3. Folland et al (2012) point out that the magnitude of the drivers we discuss in Section 3 can all be skilfully predicted in December-February winter or the winter half-year a season or more ahead. In other seasons, understanding is much less and seasonal forecasting models commensurably much less skilful. However, the AMO is known to affect UK summer atmospheric circulation and rainfall (Folland et al., 2009; Sutton and Dong, 2012) as well as spring and autumn rainfall (Sutton and Dong, 2012) and is skilfully predictable a year or more ahead using persistence. Folland et al. (2009) also suggest an influence from strong La Niňas towards wetter than normal conditions in July and August. So a major effort in studying drivers of predictability should be made for all seasons, particularly summer, when droughts can manifest themselves most severely. Whilst the winter season is most important for replenishment of water resources in the English Lowlands, intervening summers can be influential in dictating the outcomes of droughts – as was the case for the 2010 – 2012 drought, including its dramatic termination by the summer (Parry et al. 2013). In contrast, some of the most severe droughts have been associated with the combination of one or more dry winters with subsequent arid summers (e.g. in 1976, 1989). There is therefore a need to understand the drivers of both winter half-year and summer half-year deficiencies, and the likelihood of persistence between them in driving sequences of below-normal rainfall between seasons in long droughts. Folland et al (2009) showed that in summer, the summer NAO is the most prominent atmospheric circulation pattern and especially affects the English Lowlands. Its phase strongly modulates rainfall and temperature together such that both enhance drought or flood conditions. This is because high PMSL in summer, corresponding to the positive phase of the summer NAO is associated with dry, sunny and warm conditions while cyclonic conditions, associated with the negative phase, are associated with wet, dull and cooler conditions. Long droughts can also terminate at the end of summer dramatically, e.g. that of 1975-1976 (Folland, 1983).

Because many complex dynamical processes are involved, non-linear interactions may be important in creating the climatic outcome from a given combination of predictors. Only climate models can, in principle, represent these interactions as observed data are too few for reliable non–linear statistical methods. Furthermore, the climate is in any case becoming increasingly non-stationary as global temperatures increase. It used to be thought that increasing greenhouse gases would most likely be associated with a slow tendency to an increasing positive, westerly phase of the winter NAO over the UK (e.g. Gillett et al., 2003). However a recent tendency towards more negative winter Arctic and North Atlantic Oscillations casts doubt on this result (Hanna et al., 2014). Furthermore, ten dynamical models with high resolution stratospheres suggest that increasing greenhouse gases may be associated with a tendency to more winter blocking over higher northern latitudes with perhaps some increased frequency of easterly winds over northern UK in winter compared to current climate (Scaife et al., 2012). The net effect on winter English Lowlands rainfall is by no means certain, though Scaife et al find increased winter rainfall. In summer, there is more consensus that anticyclonic conditions may increase in the long-term under increased greenhouse gases in southern UK with decreased English Lowlands summer rainfall (e.g. Rowell and Jones, 2006, Folland et al, 2009). It is increasingly clear, though, that AMO fluctuations, which themselves may be influenced by anthropogenic forcing, may for decades reduce or hide this tendency or temporarily enhance it. However Arctic sea ice reductions might affect long term summer trends in hitherto unexpected ways (Belflamme et al., 2013), and become an important influence in all seasons. Despite considerable uncertainty around changes in precipitation patterns, projections for future increases in temperature for the UK are more robust. The associated increases in evapotranspiration are likely to be a further factor increasing drought severity in future.

4.4 **The way forward**

Recent developments in climate modelling (e.g. Hazeleger et al., 2010, Scaife et al., 2011, Maclachlan et al., 2014) provide the key way forward for investigating European climate mechanisms, supported by observational studies using improving and temporally expanded reanalyses. Dynamical climate models can be run in various complimentary ways. This includes running coupled ocean-atmosphere models, running their atmospheric component (AGCM) against observed lower boundary layer forcing, particularly SST and sea ice extents, and carrying out special experiments with specified forcings like observed SST patterns, including ENSO, or combinations of other forcings discussed above.

Recent research indicates that using AGCMs with specified SST and sea ice (e.g. HadISST1, Rayner et al., 2003) is a useful way forward for predictability studies though there are limitations (e.g. Chen and Schneider, 2014). This may allow estimates of UK and perhaps English Lowlands rainfall predictability through the seasonal cycle, for example using the newly improved HadISST2 data set (Titchner and Rayner, 2014). An advantage of such runs is that SST variations are realistic whereas they may not be in coupled models.

Coupled models have already shown great promise as shown by the high skill of an ensemble of retrospective December-February European forecasts from a high resolution version of the HadGEM3 coupled ocean-atmosphere climate model run for the last 20 winters (Scaife et al., 2014a). The SST predictions for this season also show considerable skill (MacLachlan et al., 2014). This work also shows that some aspects of the seasonal surface climate prediction can be further improved by basing them on forecasts of the governing atmospheric circulation pattern rather than the directly forecast surface conditions *per se*. For example, prediction of the NAO is more skilful than, say, the prediction of temperature across northern Europe but because the NAO often governs regional climate fluctuations, European winter surface climate predictions may be improved if derived from the forecast NAO (Scaife et al., 2014a), at least in some regions. Thus a good way to use dynamical seasonal climate predictions of regional UK rainfall in a hydrological context may be to combine dynamical atmospheric circulation predictions with statistical downscaling. A combination of atmospheric and coupled model approaches might be particularly valuable for studying the hitherto unknown causes of the large and persistent atmospheric circulation changes that resulted in the sudden ends of some major droughts like those of 1975-76 and 2010-2012.

The 20CR stretching back to 1871, now in an enhanced version 2 form (http://www.esrl.noaa.gov/psd/data/gridded/data.20thC_ReanV2.html) and other existing and planned reanalyses will allow new observational studies of relationships between predictors, atmospheric circulation through the depth of the troposphere and rainfall for more than the last century. Thus the late 19th century and very early 20th century is an especially interesting period for study. It included several major English Lowland drought episodes, including a long drought from

1854-1860, a major drought from 1887-1888 and the ‘Long Drought’ of 1890-1910 (Marsh et al., 2007; Todd et al., 2013). The latter was associated with several clusters of dry winters analogous to some recent multi-annual droughts. Such studies emphasise the importance of further digitizing historical rainfall data. For example, digitized UK rainfall records from paper archives would enable key datasets such as NCIC rainfall to be pushed back into well into the late 19th Century. This, coupled with the longevity of the 20CR data, would open up new possibilities for examining the climatic drivers behind these multi-annual droughts of the 19th Century. As indicated in section 4.1, a key issue in long, multi-annual droughts is the sequencing between dry winter and summer half-years. The use of long hydrometric records opens up the possibility of exploring frequency-duration relationships to examine drought persistence in a probabilistic sense, e.g. using Markov Chain models to explore dry(wet) to dry(wet) season persistence (Wilby, in preparation)

A key area for further study is improved understanding of the hydrological response to precipitation deficits during the onset, development of and recovery from, drought episodes. This study has used consistent indicators of rainfall, flow and groundwater to shed new light on temporal correlations between meteorological drought anomalies (SPI) and their response in river flow (SSI) and groundwater levels (SGI). However, this has only been evaluated at a broad scale for the English Lowlands – the temporal relationships will vary widely across the study domain, depending on aquifer properties (Bloomfield and Marchant, 2013) and catchment properties (Fleig et al., 2011; Chiverton et al. in 2015). The study highlights the need for more systematic studies of drought propagation using a combination of observational and catchment modelling approaches (e.g. as carried out for one English catchment by Peters et al., 2006, and for selected European catchments by Van Loon et al. 2012). Finally, it is important to emphasise that the manifestation of drought impacts in the English Lowlands will be heavily influenced by water management infrastructure and societal responses (e.g. the effects of surface and groundwater abstractions, reservoir operations, and the influence of societal demand during drought events). This study has examined the region at a coarse scale, but an examination of the finer catchment/aquifer scale links between climate drivers and flow/groundwater responses will require an appreciation of the moderating role these influences will have on the propagation of climate drivers through to streamfow and groundwater responses.

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Table 1. Fifteen key 13- to 26-month duration meteorological droughts across the English Lowlands, 1910 to 2012, based on NCIC gridded rainfall data.

Table 1 is ordered by drought severity, expressed as percentage of long term average rainfall. The Niňo 3.4 SST anomaly is the average for all winter half-year months during the drought.

Start month	End month	Duration (months)	Total rainfall (mm)	1961-1990 average (mm)	Deficit (mm)	% of average	Winter Nino3.4 SST anom.	Category of La Niñ Niño
May-1975	Aug-1976	16	541	898	357	60	-1.32	Strong La Niña
Aug-1920	Dec-1921	17	630	991	361	64	-0.42	Cold Neutral
Feb-1943	Jun-1944	17	662	937	276	71	-0.66	Weak La Niña
Apr-1995	Apr-1997	25	1004	1411	407	71	-0.62	Weak La Niña
Apr-1933	Nov-1934	20	829	1133	304	73	-0.83	Weak La Niña
Mar-1990	Feb-1992	24	1006	1361	354	74	0.81	Weak El Niño
Dec-1963	Feb-1965	15	639	855	215	75	-0.17	Cold Neutral
Jun-1937	Jun-1938	13	556	735	179	76	-0.25	Cold Neutral
Aug-1988	Nov-1989	16	702	924	222	76	-1.49	Strong La Niña
Feb-1962	Feb-1963	13	556	726	170	77	-0.29	Cold Neutral
Apr-2010	Mar-2012	24	1050	1361	311	77	-1.14	Strong La Niña
Apr-1928	Sep-1929	18	782	1006	224	78	-0.03	Cold Neutral
Aug-1972	May-1974	22	995	1255	260	79	-0.07	Cold Neutral
Nov-2004	Apr-2006	18	810	1025	215	79	-0.02	Cold Neutral
Aug-1947	Sep-1949	26	1181	1478	296	80	-0.19	Cold Neutral

Table 2 Top 20 winter half-year La Niňas and English Lowlands rainfall since 1910-1911, indicating whether these correspond to the meteorological droughts in Table 1 (as described in Sect 2.2)

WINTER HALF YEAR	La Nina SST anomaly,°C, (from 1961-90)	Table 1 Meteorological Drought lasting 5-6 months in given winter	Rainfall anomaly mm/month
1988-1989	-1.87	YES	-15.2
1973-1974	-1.82	YES	-9.3
2007-2008	-1.56		1.7
1942-1943	-1.46		2.3
1999-2000	-1.43		-6.8
2010-2011	-1.42	YES	-10.5
1998-1999	-1.39		-15.2
1975-1976	-1.32	YES	-26.0
1970-1971	-1.25		4.2
1916-1917	-1.20		4.5
1949-1950	-1.10		9.3
1984-1985	-1.09		-0.2
1933-1934	-1.05	YES	-19.7
1955-1956	-1.02		-5.7
1924-1925	-0.89		8.7
1938-1939	-0.88		14.7
2011-2012	-0.86	YES	-18.9
1995-1996	-0.85	YES	-10.5
1983-1984	-0.71		1.0
1910-1911	-0.71		8.3

Table 3. Summary of remote drivers of English Lowlands rainfall.

Only the influence on English Lowlands climate are summarised; effects elsewhere in UK may be larger or different. Conditions that favour drier winters are highlighted in yellow

Climate driver	Effect on English Lowlands winter half-year precipitation and temperature
ENSO	El Niño tends to give somewhat wetter conditions than normal, while La Niño tends to give somewhat drier conditions than normal. There are intra-seasonal variations in these effects (Supplementary Info S1)
North Atlantic tripole SST anomaly	A negative North Atlantic SST tripole index in May weakly favours dry conditions in English Lowlands in the following winter half year. A positive index marginally favours wetter than normal conditions.
QBO	The QBO has only a small direct influence. A westerly QBO gives no significant rainfall signal, while a strong easterly QBO tends to give modestly wetter than average conditions. However, the rather strong effect of more extreme QBO phases on North Atlantic atmospheric circulation might modulate influences of other factors.
Major tropical volcanic eruptions	Major tropical volcanic eruptions are uncommon. They tend to force the positive westerly phase of the NAO in winter associated with wetter than normal conditions in northern Scotland and slightly drier than normal conditions much further south, including the English Lowlands.
Solar effects	Cyclonic anomalies associated near or just after solar maxima may be associated with a tendency to wetter than normal conditions
AMO	A negative NAO tends to occur when the AMO is positive and a positive NAO when the AMO is negative. However, neither phase of the AMO provides a rainfall signal for the English Lowlands. Differing intra-seasonal influences and interactions with other forcing factors cannot be ruled out.

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Figure Captions

Fig 1. Map of the English Lowlands study region (bold line indicates boundary), the river Thames (blue) and its catchment above the Kingston gauging station (red) and the location of the Rockley borehole (red). For context, the map also shows the location of London, major aquifers (light grey) and upland areas over 200m (dark grey)

Fig 2. Example of a meteorological drought, April 2010 to March 2012

Fig 3. Correlations of designated district average rainfalls with 5 x 5 km gridded rainfall data elsewhere in UK for winter and summer half-years of droughts identified in this paper. N is the calculated equivalent number of independent rainfall stations across the UK in Table 1 droughts, a measure of spatial rainfall anomaly variability in the droughts, where rainfall anomalies are differences from their long-term means.

Fig 4a. Heatmap of the correlation between lagged English Lowlands river flow SSI over a one-month timescale and English Lowlands precipitation as SPI over 1–24 months, with maximum correlation highlighted with black circle.

Fig 4b.. Heatmap of the correlation between lagged English Lowlands groundwater level SGI over a one-month timescale and English Lowlands precipitation as SPI over 1–24 months, with maximum correlation highlighted with black circle.

Fig 5. SPI, SSI and SGI for regional English Lowlands series, where the first three time series are SPI based on the English Lowlands precipitation time series, with SPI 3 month rainfall accumulation, SPI 6 month rainfall accumulation and SPI 12 month rainfall accumulation; the latter two are SSI for the English Lowlands regional river flow series and SGI for the English Lowlands groundwater level time series.

Fig 6. SPI, SSI and SGI series for the Thames, where the first three are based on the Thames catchment rainfall time series, with SPI 3 month accumulation, SPI 6 month accumulation and SPI 12 month accumulation; the latter two are the SSI series for the Thames river flow at Kingston and the SGI series for the Rockley groundwater level series.

Fig 7a. Composite global SST anomalies from 1961-1990, winter half-year, over 1901-2013 when Nino 3.4 anomalies <-1.0^{o}C

Fig 7b. Top panels: Global PMSL anomalies (hPa) from the 20th Century reanalysis averaged over winter half-year for La Niñas measured by SST <-1 standard deviation over Nino 3.4, corresponding to a 1961-1990 SST anomaly <-0.92^{o}C, for two independent epochs 1876-1950 (left) and 1951-2009 (right). The standard deviation is for 1951-2010. Central panels (left): global storminess anomalies, 1951-2013 measured by anomalies of 2-7 day band pass variance of 500hPa height (dm^2), (right) west European rainfall anomalies (mm/month) 1901-2011 for La Niňas for winter half-year. Bottom panels (left): as top right panel for moderate El Niños (anomalies of 0.92^{o}C <Nino 3.4 <1.5^{o}C) (right) as central right panel but for moderate El Niňos. Dark colours are locally significant at the 5% level. Light colours on global maps only (all diagrams) are included show the patterns more clearly but are not significant. Rainfall from the Mitchell and Jones (2005) 0.5^{o} x0.5^{o} degree data set, as it is for Figs. 9-12.

Fig 7c Cumulative distributions of English Lowlands rainfall, 1901-2014, expressed as a percentage of the 1961-90 average, for (a) La Nina and (b) El Nino conditions excluding extreme El Ninos, as described in the text

Fig 8. (Top left) Global PMSL anomalies (hPa) in winter half-year for a tripole SST index <-1 SD; (Top right) >1 SD in the previous May. (Bottom left) Rainfall anomalies in winter half-year (mm/month) over UK and nearby Europe for tripole SST index <-1 SD. (Bottom right) for >1 SD. Areas significant at the 5% level are darkly coloured. Tripole SD calculated for May 1949-2008. PMSL comes from the NCEP Reanalysis.

Fig 9. (Top left) Near global PMSL anomalies (hPa) in winter half-year for most easterly QBO 15% of 30hPa equatorial stratospheric winds (1953-1954 to 2012-2013). (Top right) Rainfall anomalies for the top 15% most easterly of all equatorial winds. (Bottom left) As top left but for the 15% most westerly QBO winds. (Bottom right) As top right, but for the 15% most westerly winds. Areas significant at the 5% level are dark coloured. PMSL is from the NCEP Reanalysis.

Fig 10. (Left) Near global PMSL anomalies (hPa) in winter half year for TSI values in the highest 20% of the its winter half year distribution over 1948-2011. Earlier years not used as solar cycle mostly varied at an averaged reduced level of total solar radiation. (Right) Rainfall anomalies (mm/month) over UK and nearby Europe. Areas significant at the 5% level are darker coloured.

Fig 11. (Top left) Near global PMSL anomalies (hPa) in winter half year for monthly AMO index values <-1SD calculated over 1871-2013. (Top right) rainfall anomalies (mm/month) for AMO index values <-1SD. (Bottom left) Near global PMSL anomalies for AMO index values >1SD (Bottom right) Rainfall anomalies (mm/month) for AMO Index values >1SD. Areas significant at the 5% level are darker coloured. PMSL is from the 20CR

Fig 12a. Box plots of English Lowland response variables for the October to March winter half year (English Lowlands areal rainfall and total flow), for low (<-0.5 SD) and high (>0.5 SD) values of different drivers (Niño 3.4, IPO, TSI, May SST tripole, AMO, stratospheric aerosol loadings and QBO).

Fig 12b. Box plots of English Lowland response variables for the October to March winter half year (SSI flow, SGI Groundwater andthree accumulation periods for the SPI), for low (<-0.5 SD) and high (>0.5 SD) values of different drivers (Niño 3.4, IPO, TSI, May SST tripole, AMO, stratospheric aerosol loadings and QBO).

FIG 1

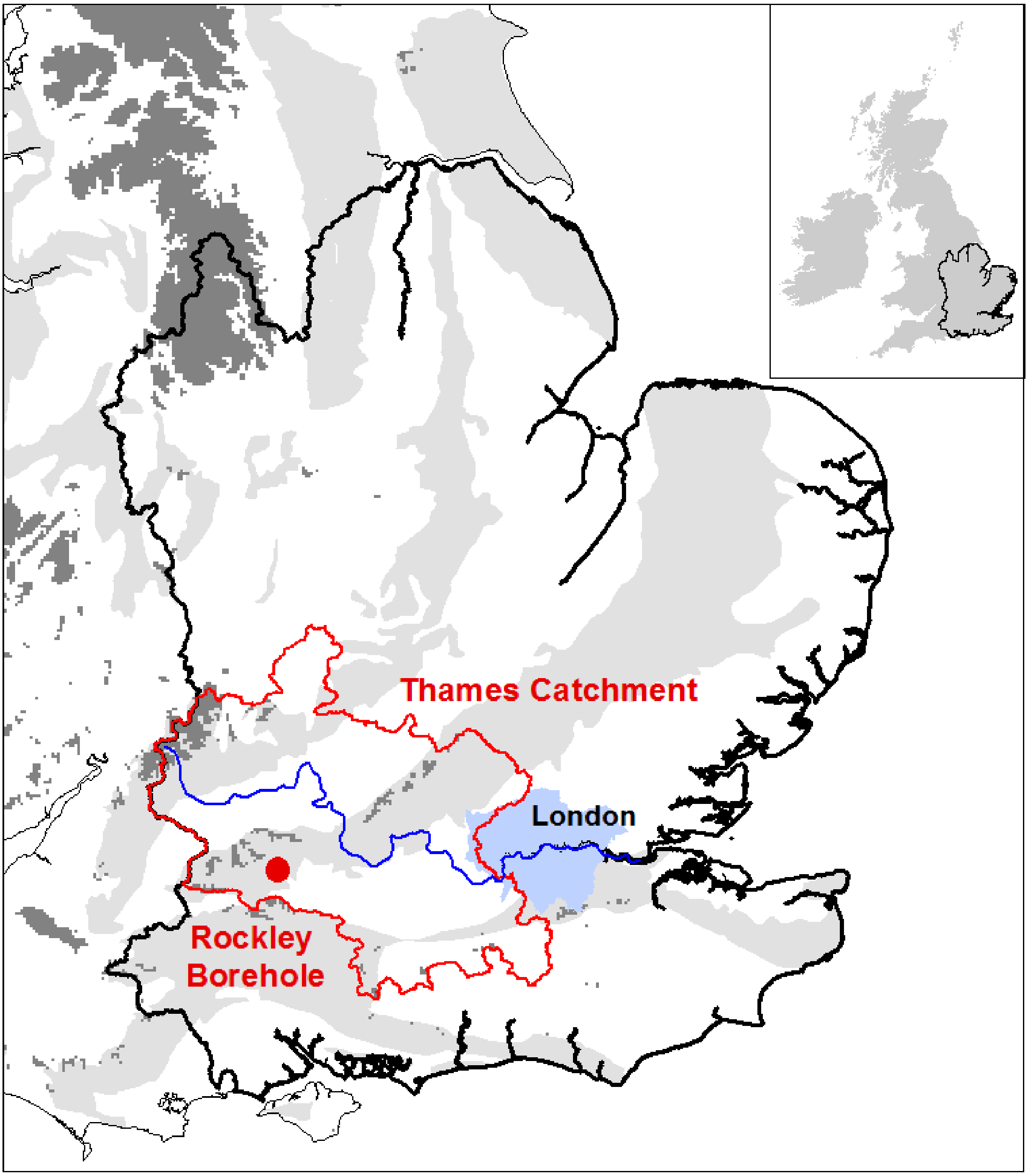


FIG 2

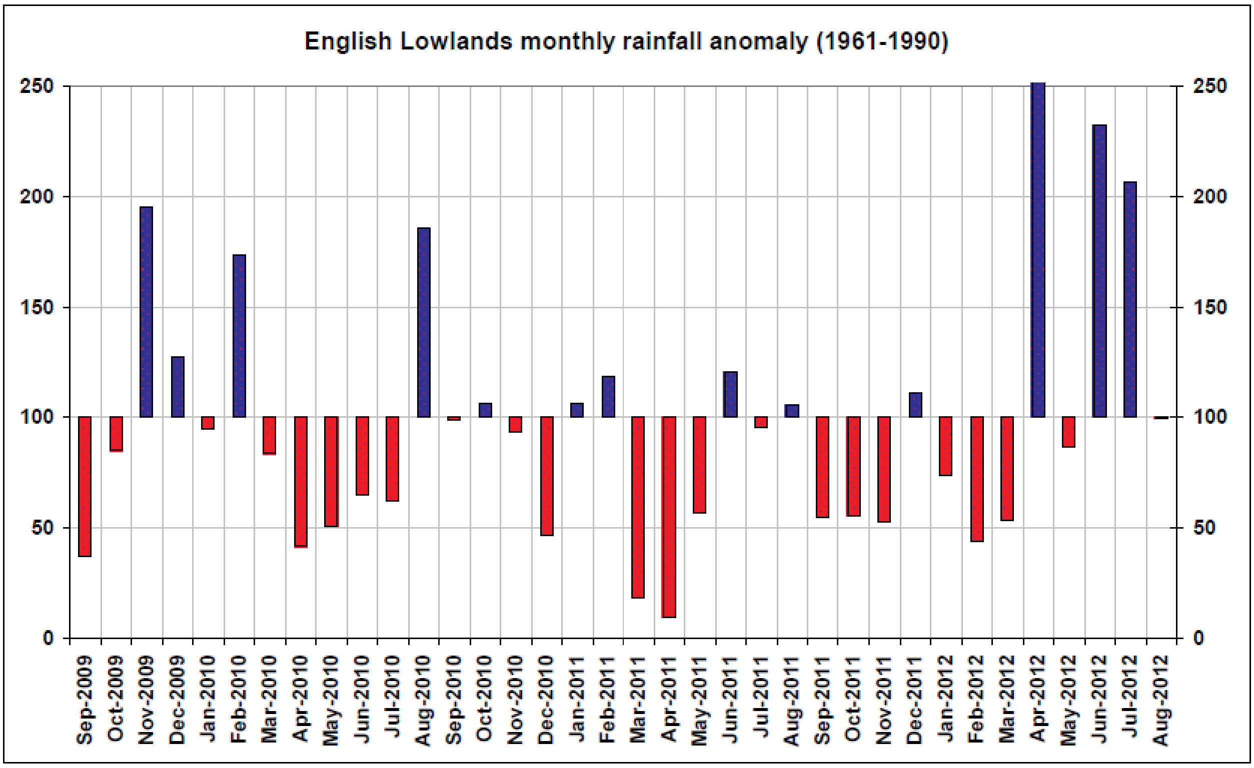

English Lowlands monthly rainfall anomaly (1961-1990)
250
200
150
100
50
0
Sep-2009
Oct-2009
Nov-2009
Dec-2009
Jan-2010
Feb-2010
Mar-2010
Apr-2010
May-2010
Jun-2010
Jul-2010
Aug-2010
Sep-2010
Oct-2010
Nov-2010
Dec-2010
Jan-2011
Feb-2011
Mar-2011
Apr-2011
May-2011
Jun-2011
Jul-2011
Aug-2011
Sep-2011
Oct-2011
Nov-2011
Dec-2011
Jan-2012
Feb-2012
Mar-2012
Apr-2012
May-2012
Jun-2012
Jul-2012
Aug-2012

FIG 3

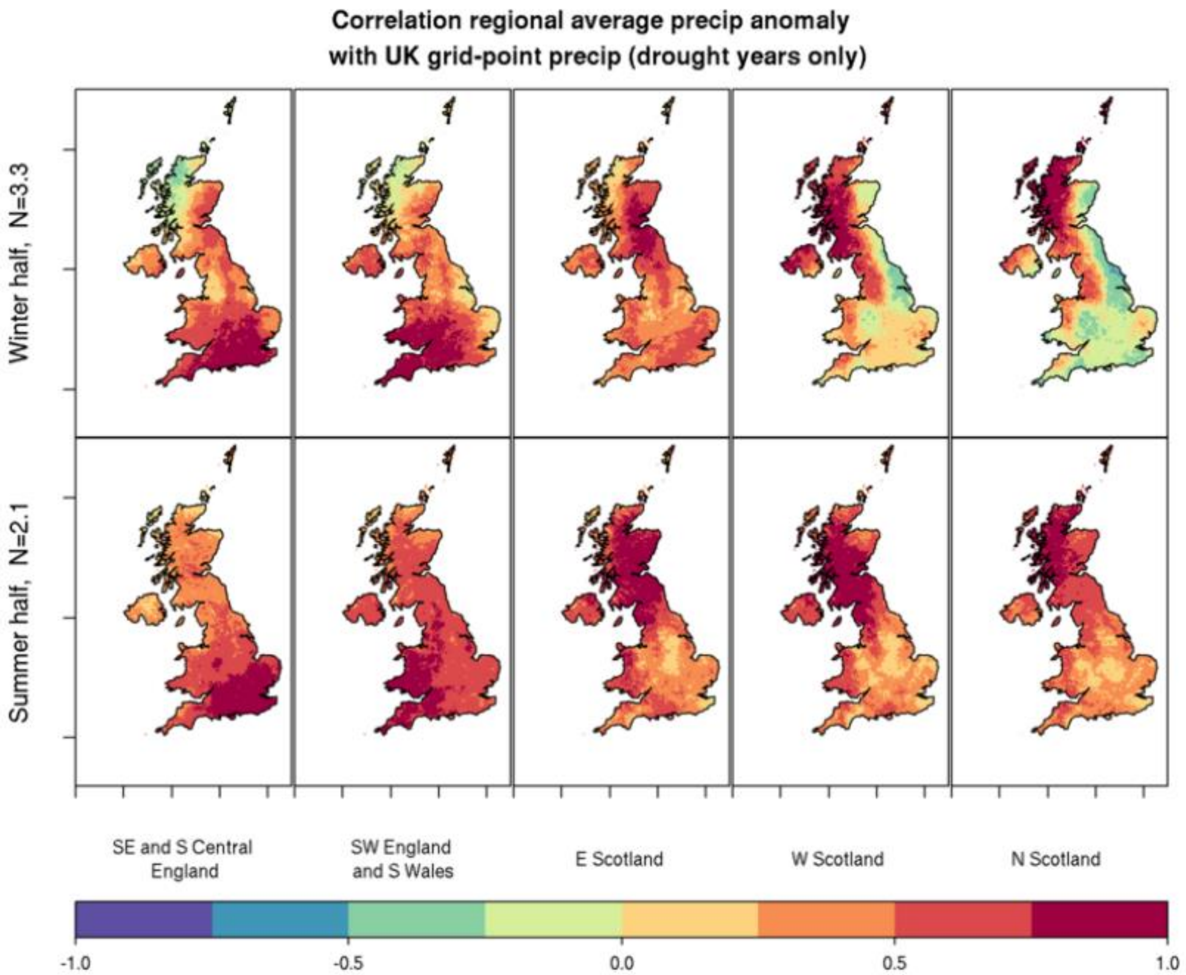
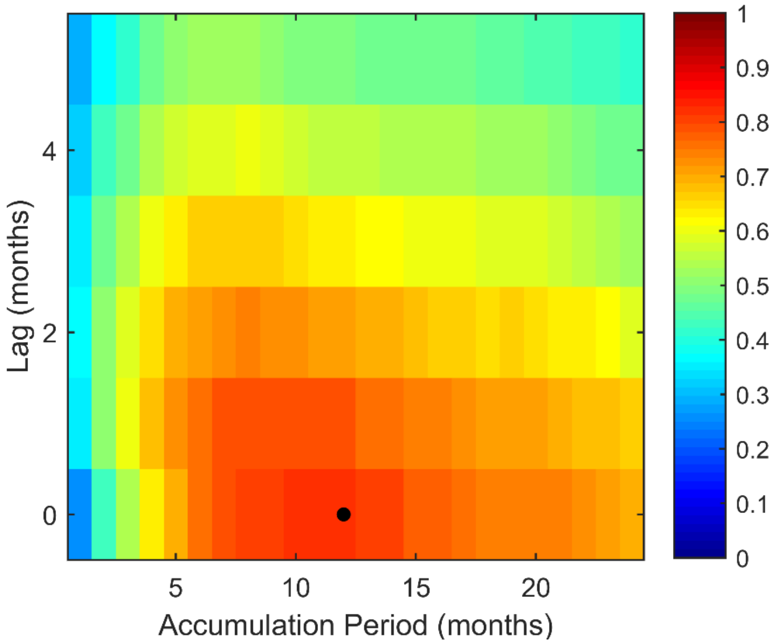

Correlation regional average precip anomaly
with UK grid-point precip (drought years only)
Winter half, N=3.3
Summer half, N=2.1
SE and S Central
England
SW England
and S Wales
E Scotland
W Scotland
N Scotland
-1.0
-0.5
0.0
0.5
1.0

FIG 4a and FIG 4b



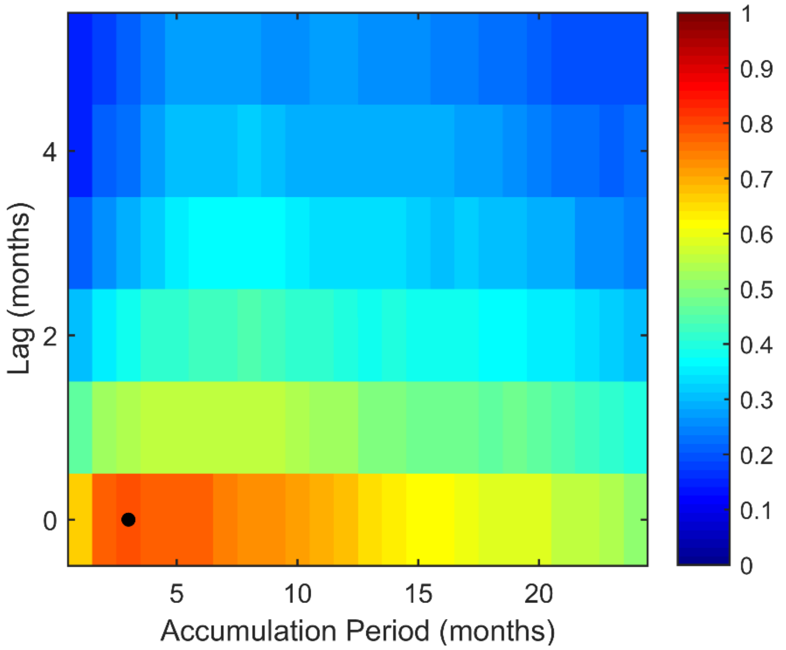
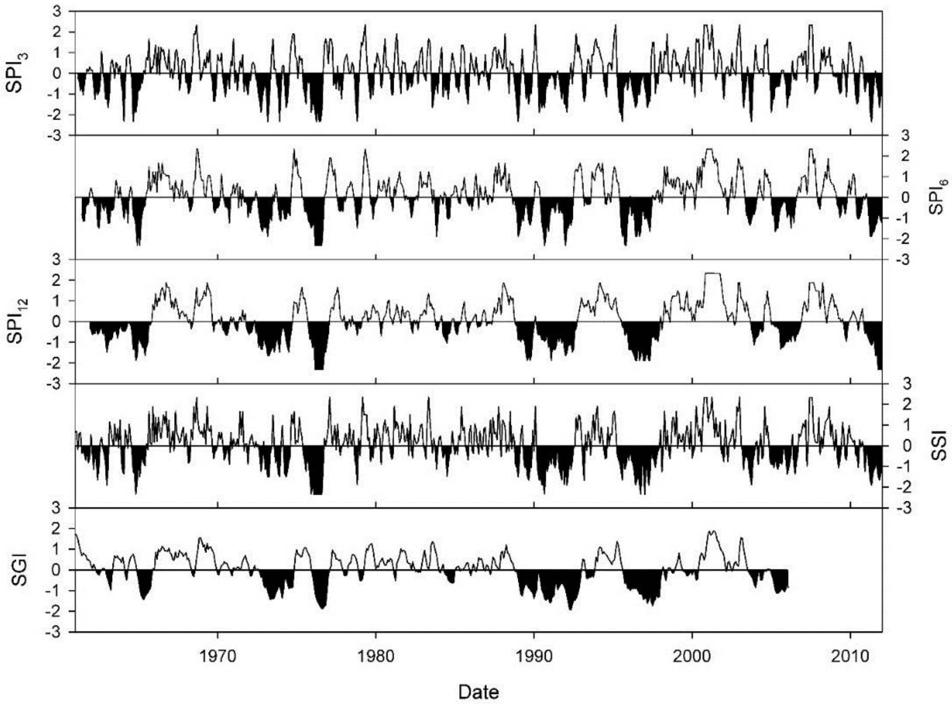


FIG 5



SPI3
SPI6
SPI12
SSI
SGI
3
2
1
0
-1
-2
-3
1970
1980
1990
2000
2010
Date

FIG 6

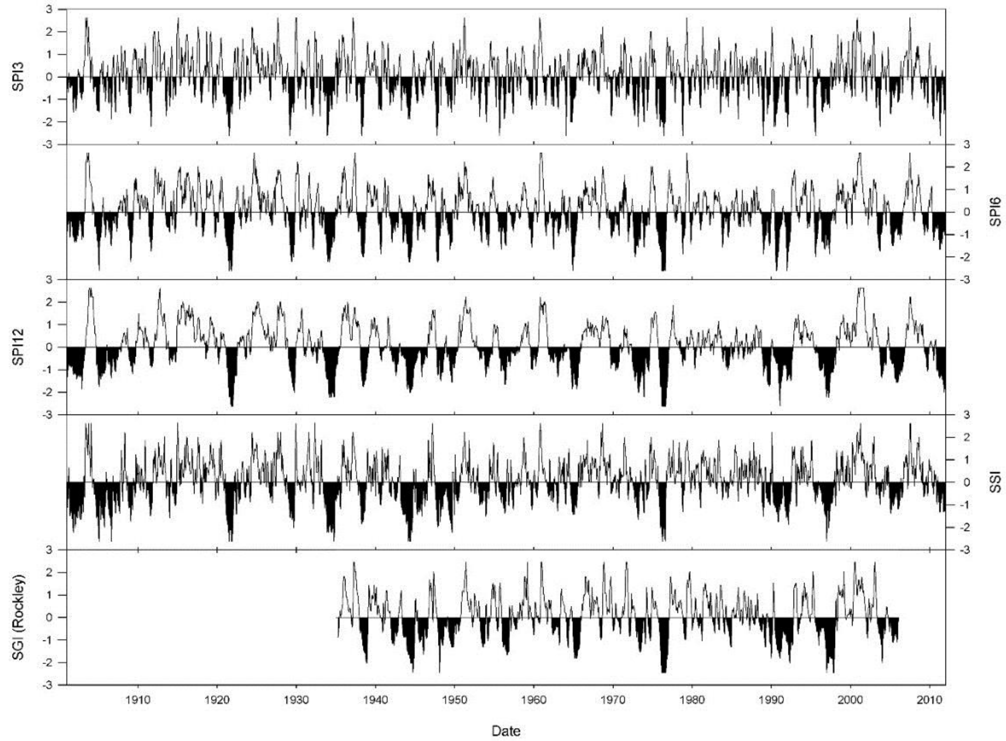
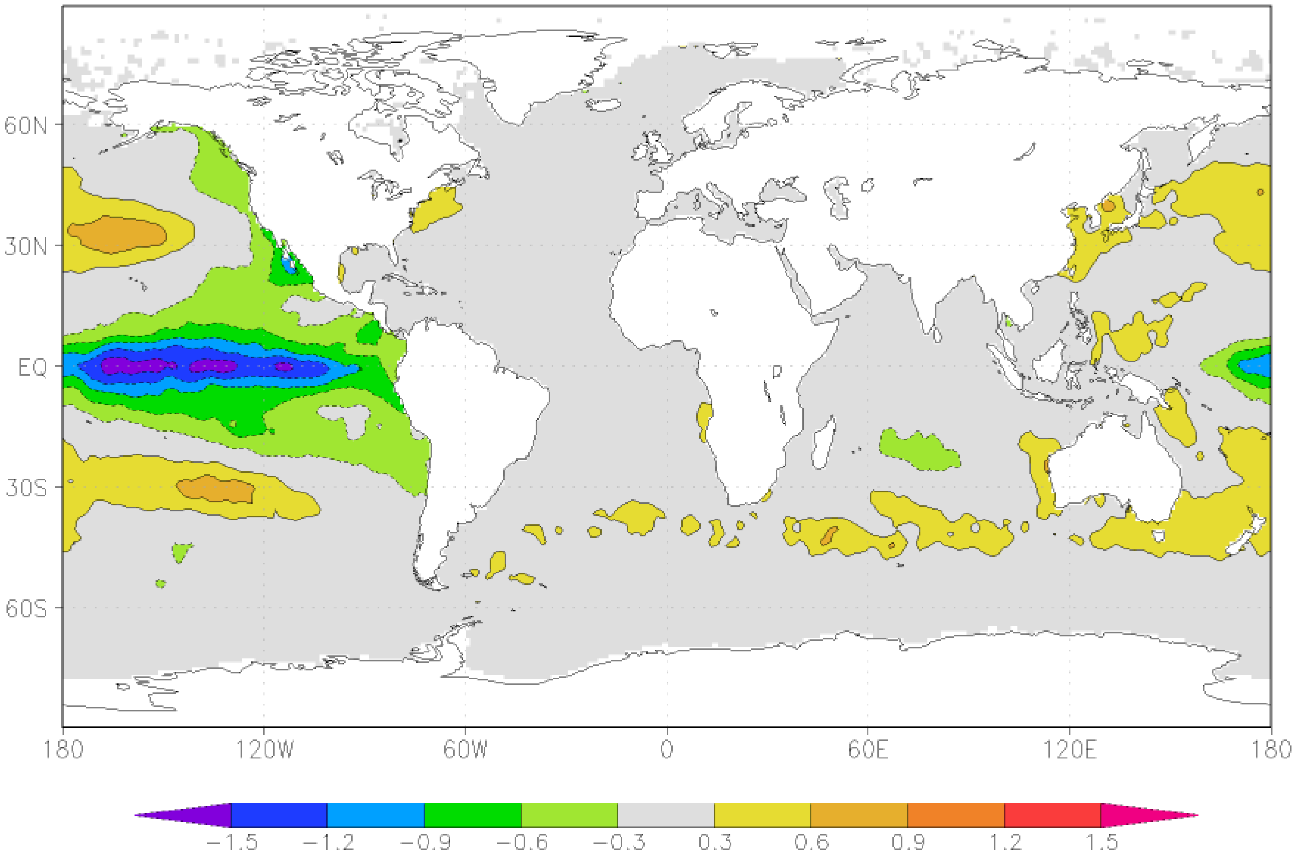

SPI3
SPI6
SPI12
SSI
SGI (Rockley)
1910
1920
1930
1940
1950
1960
1970
1980
1990
2000
2010
Date

FIG 7a



60N
30N
EQ
30S
60S
180
120W
60W
0
60E
120E
180
−1.5
−1.2
−0.9
−0.6
−0.3
0.3
0.6
0.9
1.2
1.5

FIG 7b

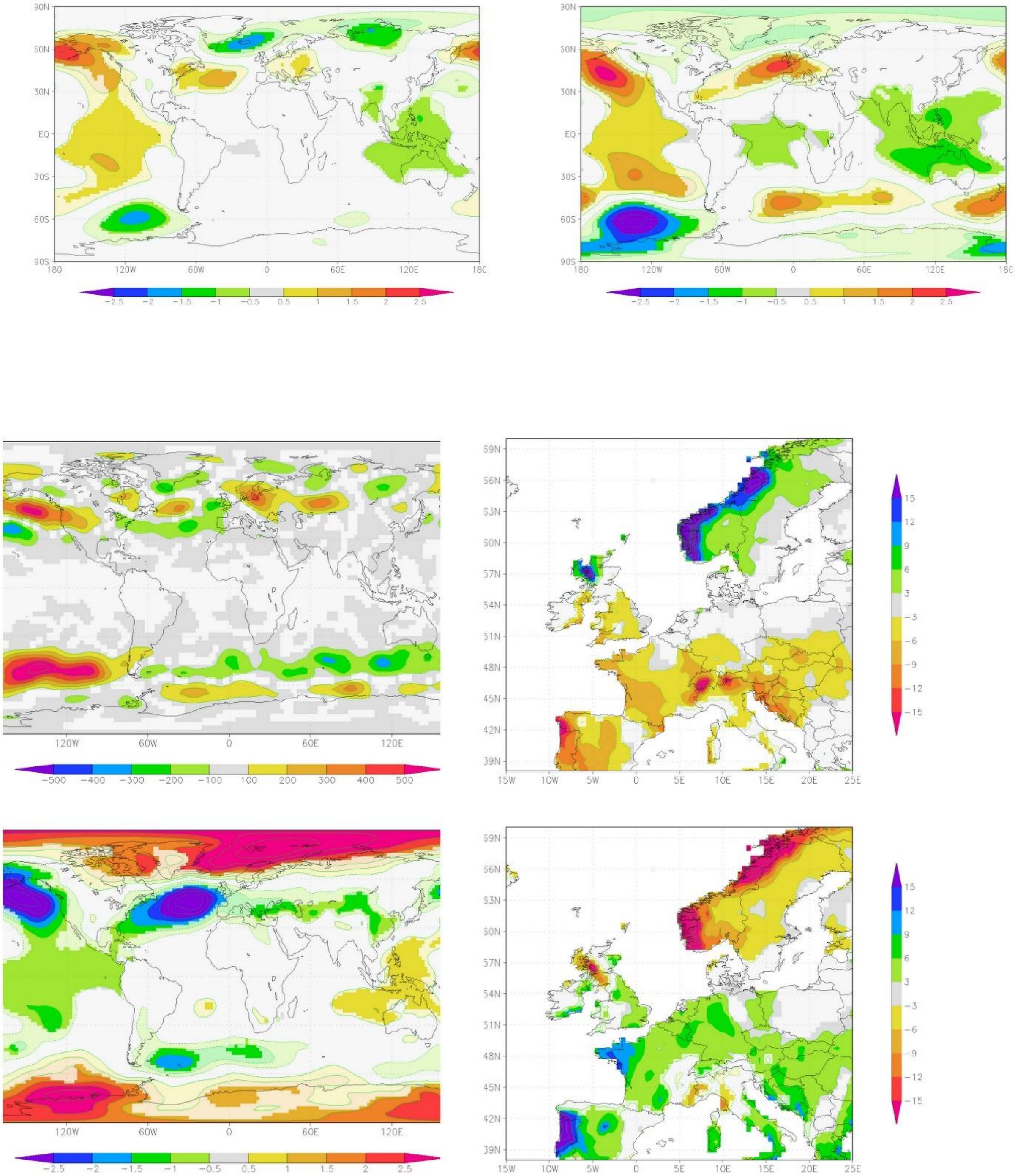


FIG 7c

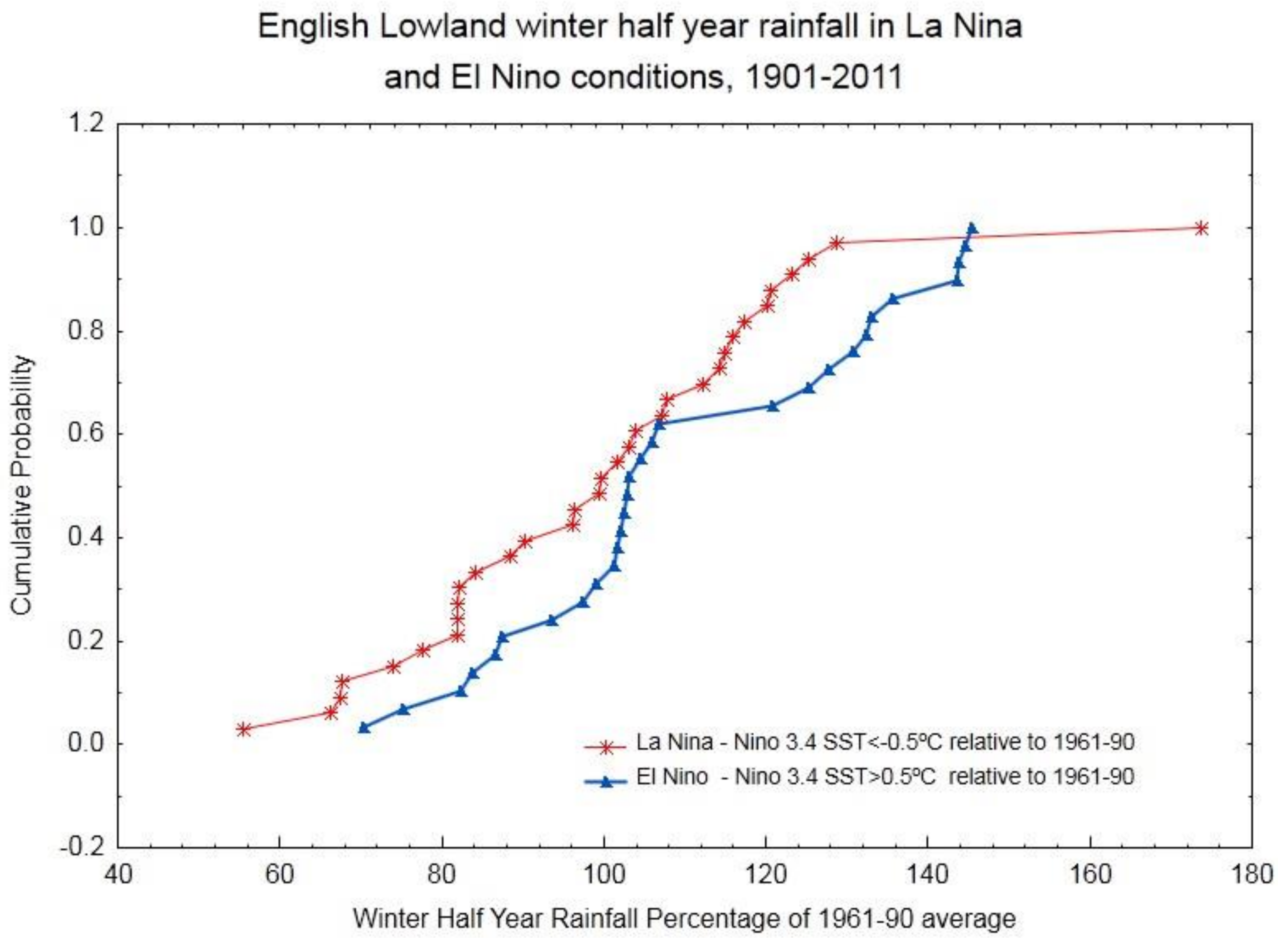

English Lowland winter half year rainfall in La Nina
and El Nino conditions, 1901-2011
Cumulative Probability
Winter Half Year Rainfall Percentage of 1961-90 average
La Nina - Nino 3.4 SST<-0.5ºC relative to 1961-90
El Nino - Nino 3.4 SST>0.5ºC relative to 1961-90
1.2
1.0
0.8
0.6
0.4
0.2
0.0
-0.2
40
60
80
100
120
140
160
180

FIG 8

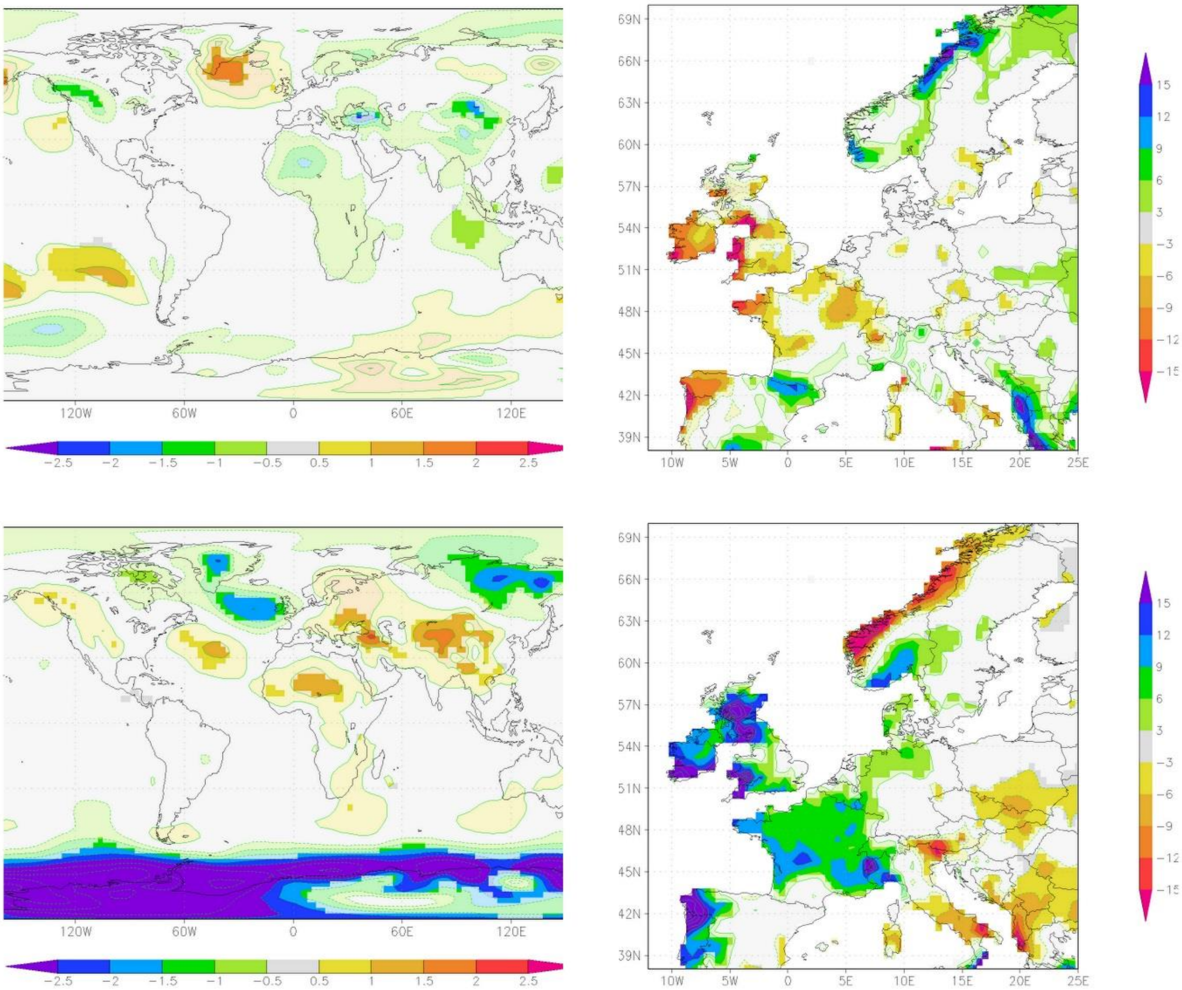

120W
60W
0
60E
120E
−2.5 −2 −1.5 −1 −0.5 0.5 1 1.5 2 2.5
69N
66N
63N
60N
57N
54N
51N
48N
45N
42N
39N
10W
5W
0
5E
10E
15E
20E
25E
15
12
9
6
3
−3
−6
−9
−12
−15

FIG 9

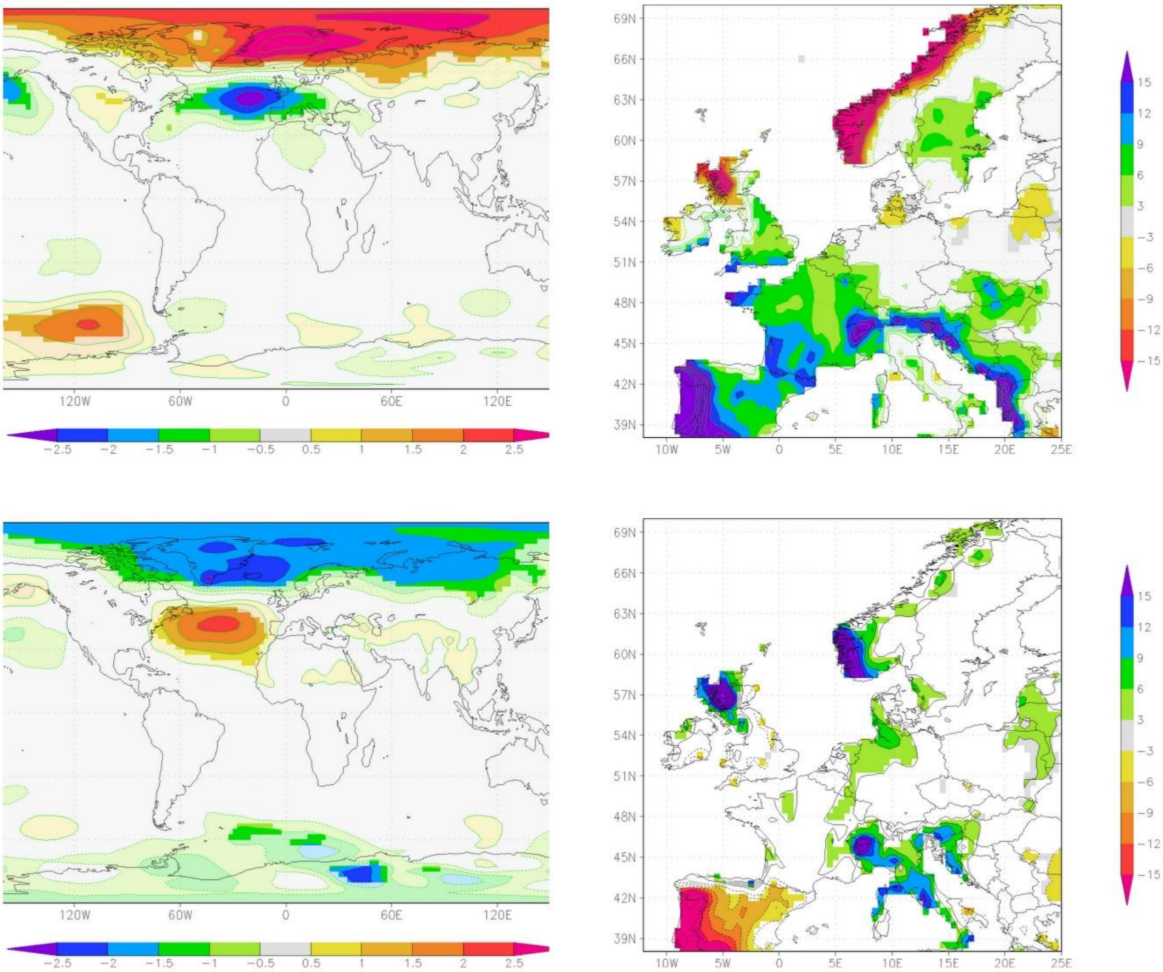

120W 60W 0 60E 120E
−2.5 −2 −1.5 −1 −0.5 0.5 1 1.5 2 2.5
69N 66N 63N 60N 57N 54N 51N 48N 45N 42N 39N
10W 5W 0 5E 10E 15E 20E 25E
15 12 9 6 3 −3 −6 −9 −12 −15

FIG 10

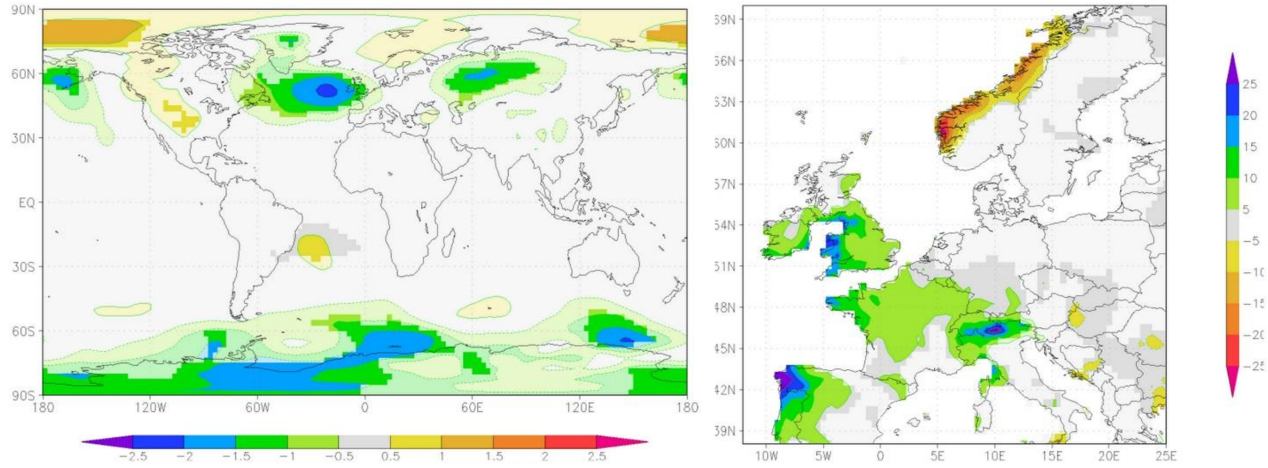

90N
60N
30N
EQ
30S
60S
90S
180
120W
60W
0
60E
120E
180
−2.5 −2 −1.5 −1 −0.5 0.5 1 1.5 2 2.5
10W
5W
0
5E
10E
15E
20E
25E
25
20
15
10
5
−5

Fig 11

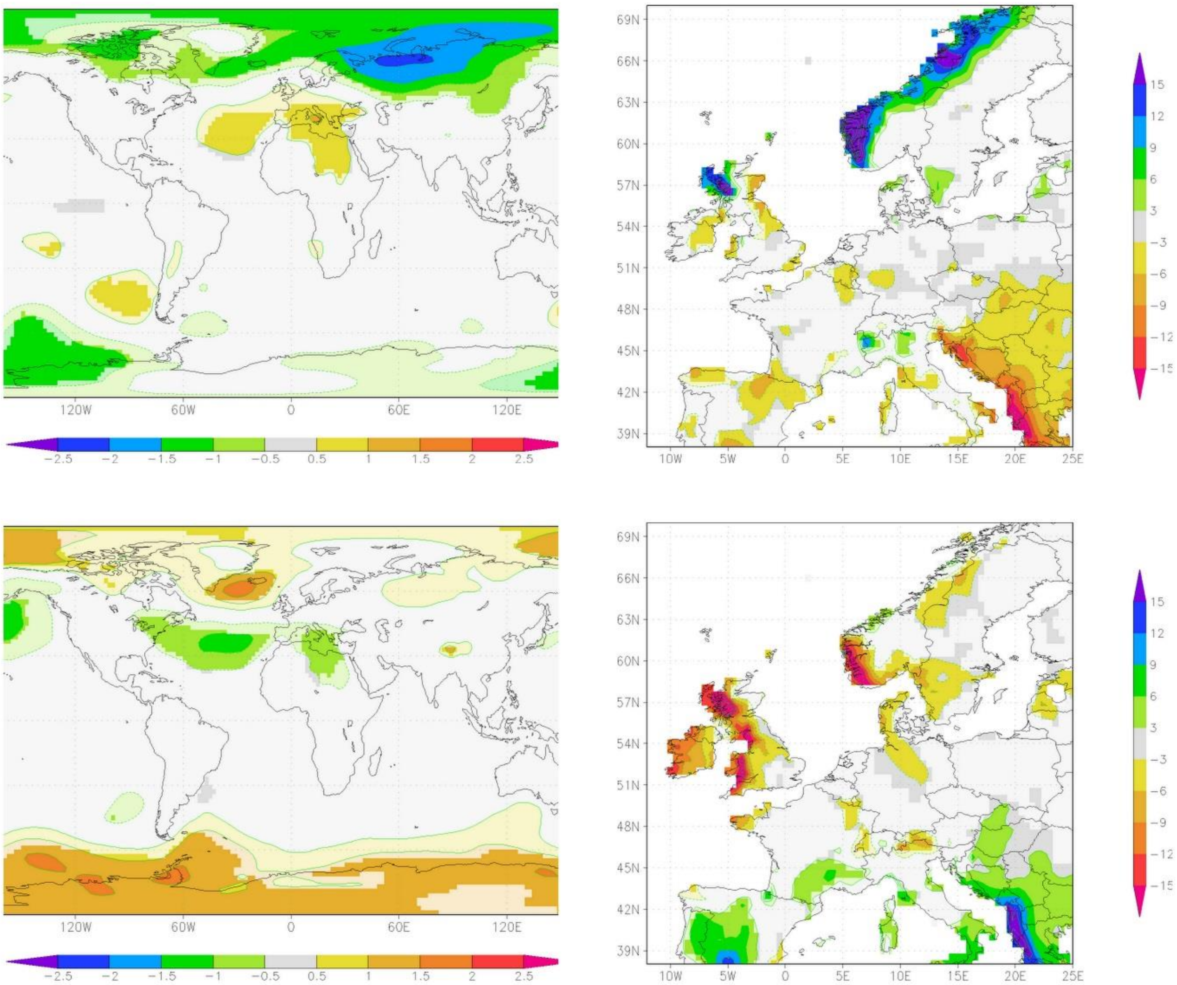

120W
60W
0
60E
120E
-2.5
-2
-1.5
-1
-0.5
0.5
1
1.5
2
2.5
69N
66N
63N
60N
57N
54N
51N
48N
45N
42N
39N
10W
5W
0
5E
10E
15E
20E
25E
15
12
9
6
3
-3
-6
-9
-12
-15

FIG 12a

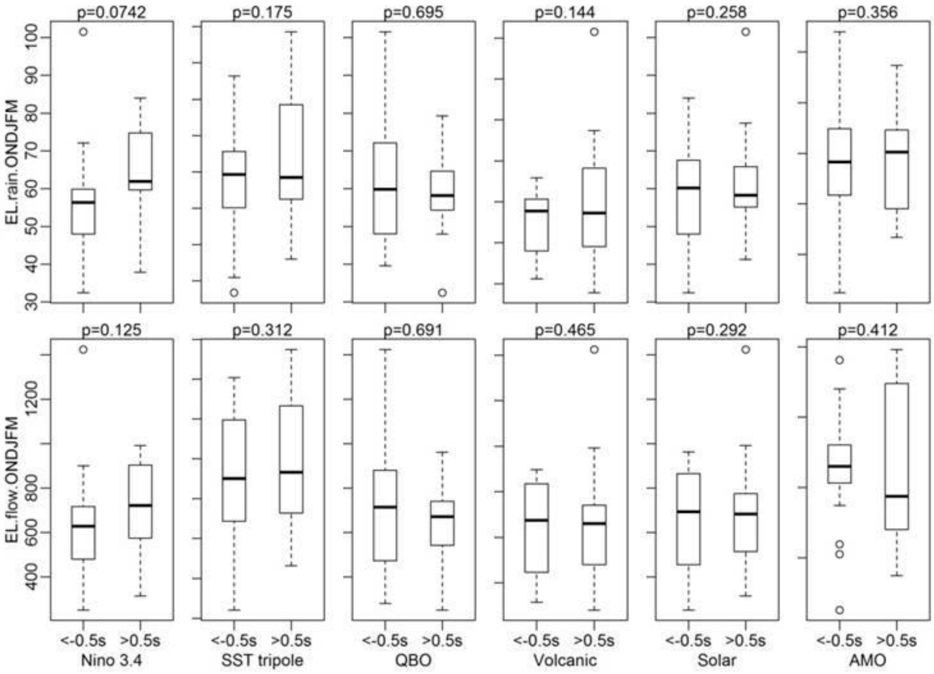
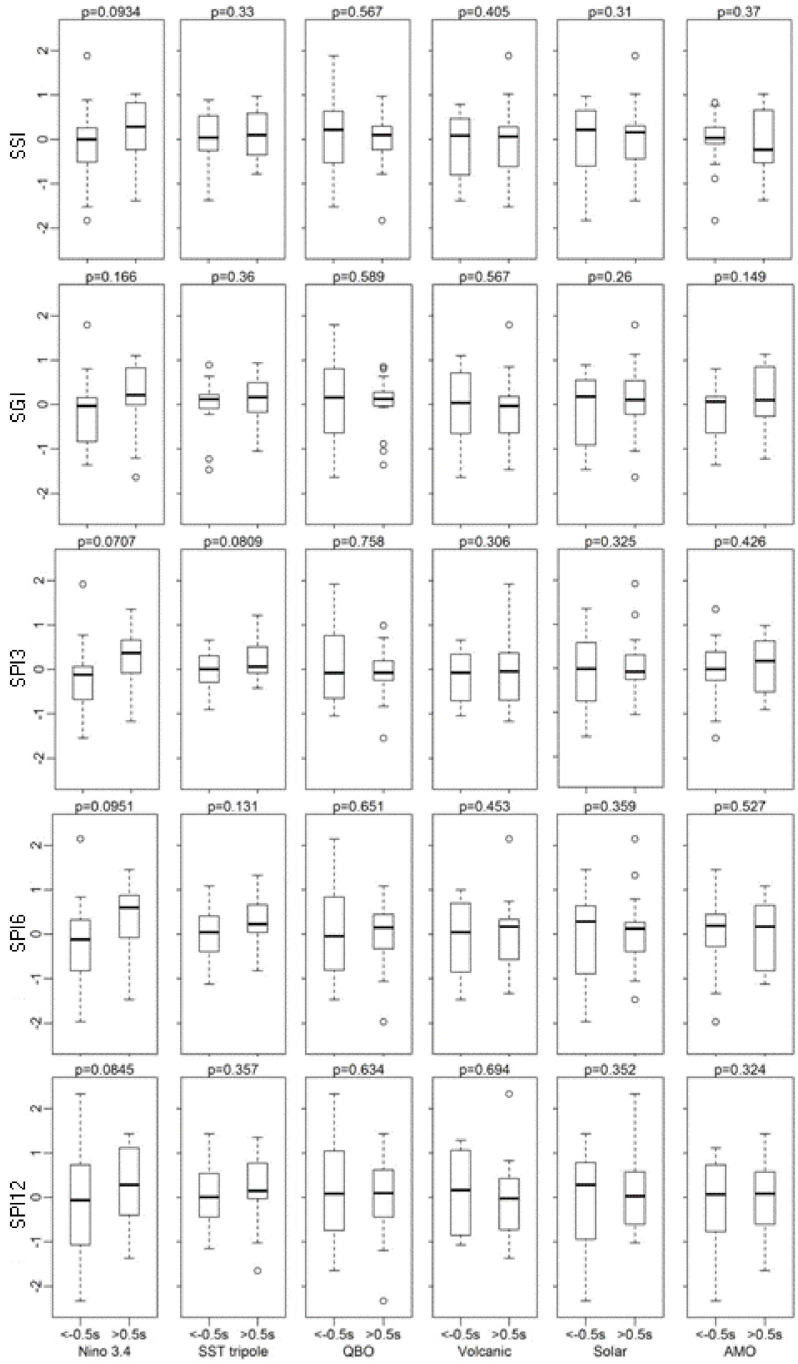

p=0.0742
p=0.175
p=0.695
p=0.144
p=0.258
p=0.356
EL.rain.ONDJFM
30
40
50
60
70
80
90
100
p=0.125
p=0.312
p=0.691
p=0.465
p=0.292
p=0.412
EL.flow.ONDJFM
400
600
800
1200
<-0.5s >0.5s
Nino 3.4
<-0.5s >0.5s
SST tripole
<-0.5s >0.5s
QBO
<-0.5s >0.5s
Volcanic
<-0.5s >0.5s
Solar
<-0.5s >0.5s
AMO

FIG 12b



p=0.0934
p=0.33
p=0.567
p=0.405
p=0.31
p=0.37
p=0.166
p=0.36
p=0.589
p=0.567
p=0.26
p=0.149
p=0.0707
p=0.0809
p=0.758
p=0.306
p=0.325
p=0.426
p=0.0951
p=0.131
p=0.651
p=0.453
p=0.359
p=0.527
p=0.0845
p=0.357
p=0.634
p=0.694
p=0.352
p=0.324
SSI
SGI
SPI3
SPI6
SPI12
<-0.5s >0.5s
Nino 3.4
SST tripole
QBO
Volcanic
Solar
AMO