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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 very 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) and a Standardized 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 Multidecadal 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.

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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 area, hereafter referred to as the English Lowlands (Fig. 1), is the driest part of the UK. It has a relatively low annual average rainfall (a 1961–1990 areal average of 680 mm, with < 600 mm being common in the east of the region). The English Lowlands have the greatest density of population, intensive agriculture and commercial enterprise in the UK, and 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).

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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 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 dry 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 were 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 (Kingston et al., 2013). These studies have all 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. 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, no attempts have been made to systematically link hydrological droughts anywhere in the UK to the known range of physical drivers of variations in atmospheric circulation that affect the UK. Thus no studies have addressed links with 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.

In summary, 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 Sect. 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, 1975/76, 1988–1992, 1995–1997 and the early 2000s. More recently, Bloomfield and Marchant (2013) developed a ground-

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water 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 for when we consider 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 km × 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

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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 km × 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 and 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, was used to provide a longer temporal coverage, alongside the NRFA catchment monthly rainfall series for that catchment. These series are available from 1883, but in the present study the post-1900 data is displayed to give coverage comparable to the English Lowlands series. 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 to remove the influence of abstractions. 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

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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. Figure 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). 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 conducted an analysis to examine how spatially coherent these major long droughts are relative to the rest of the UK. Rainfall tends to be influenced differently in northwest Britain when the English Lowlands suffer drought. To show 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 based on the 15 drought periods listed in Table 1. Although summer is not a focus of the paper, contrasts between winter and summer correlations during long period droughts are clear from Fig. 3. Generally, there is a greater contrast between southeast UK and northwest UK for winter half year than summer half year rainfall. Thus, 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

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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 is 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,

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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 SGI was also applied to the English Lowlands regional river flow series. 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, no accumulation was applied in the application of the SGI.

SGI was 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 SGI) 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.

Figure 5 shows, for the English Lowlands, SPI rainfall series for several accumulation periods and the SGI river flow and groundwater series. Figure 6 shows the SPI and SGI series for the long Thames record, the English Lowlands rainfall series and the Rockley borehole. Both figures demonstrate good agreement between the meteorological droughts and associated river flow and groundwater droughts – with some expected lags for the onset of given hydrological drought events, demonstrating the propagation between the meteorological and groundwater droughts in particular. Figure 6 also shows very good agreement between the severity of the major rainfall droughts iden-

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tified 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 anomalies in this period is 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.

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A necessary first-step in linking driving factors with rainfall anomalies is in considering their influence on surface pressure. Importantly, English Lowlands rainfall anomalies on seasonal time scales are relatively highly linearly correlated with the simultaneous pressure (pressure at mean sea-level, PMSL) anomaly over the English Lowlands. Averaged over the 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 1901–2011 (61 % of explained rainfall variance), or $-21\,\mathrm{mm\,hPa^{-1}}$ 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 Lowland England 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 some other 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 then current 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).

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Recently, a much higher level of real-time forecast skill for the NAO has been demonstrated by Scaife et al. (2014a) for core winter months of December–February for UK and Europe using the latest climate model Glosea 5 (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 indeed needed for further understanding of the full 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 $0.5^{\circ} \times 0.5^{\circ}$ rainfall data of Mitchell and Jones (2005), PMSL data of Allan and Ansell (2006), 300 hPa and PMSL data from the Twentieth Century Reanalysis (20CR) (Compo et al., 2011), the NCEP Reanalysis (Kalnay et al., 1996) and HadISST 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 the annual total solar irradiance up to 1978 from Prather et al. (2014), interpolated to monthly values, with measured monthly values from 1979 (Fröhlich, in press), May North Atlantic SST Tripole data (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.

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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, extra-tropical 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). 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 Niñ a (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 pattern associated with La Niña events where SST averaged over the Niño 3.4 region (120–170° W, 5° N–5° S) is $\leq -1.0\,^{\circ}$C below the 1961–1990 average. SST values averaging $\geq 1.0\,^{\circ}$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 h for the last 130 years using an ensemble of over 50 different slightly differing analyses. Figure 7b, top panel, shows mean PMSL anomalies (from 1961–1990) for La Niñas where Nino 3.4 region SST anomalies are $< -0.92\,^{\circ}$C for two independent epochs 1876–1950 and 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.

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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^{-1} (30 mm in the winter half year as a whole) while northwest Scotland by contrast has significant slight to moderate wet anomalies exceeding 10 mm month^{-1}. The average PMSL anomaly over the English Lowlands in 1951–2009 of 1.8 hPa in Fig. 7b corresponds to about a 38 mm rainfall deficit, 11 % of the 1961–1990 winter half year average of 348 mm. The average effect is thus modest, as with all other individual climatic influences, though individual La Niñ a 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 Supplement Fig. S1 for each winter half year month. Figure 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 °C < Niño 3.4 SST anomaly < 1.5 °C). For the relatively uncommon extreme El Niños, PMSL (Toniazzo and Scaife, 2006) and rainfall patterns change over the UK and Lowland England (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 °C), seven "neutral" conditions (anomalies between ±0.5 °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 °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

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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 −0.39SDs. 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 2000/01, the wettest winter half year in this record at 43 mm month^{-1} 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).

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

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a positive NAO displaced slightly southwards, and the negative index with a negative NAO (Fig. 8a and 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 and 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 30 hPa 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).

Figure 9 illustrates global PMSL and rainfall anomalies over UK and nearby Europe associated with strong easterly and westerly QBO winds at 30 hPa 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). 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.

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 showing that ultraviolet solar radiation variations associated with the 11 year solar cycle of total solar irradiance (TSI) modulates the Arctic Oscillation and NAO and thus winter blocking over UK through stratospheric–tropospheric interactions. 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. Scaife

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et al. (2013) showed that solar modulation of the NAO feeds back onto the North Atlantic SST Tripole. This in turn influences atmospheric circulation to create a maximum winter atmospheric circulation response 1–4 years after solar maximum (westerly phase of the NAO) and solar minimum (easterly phase of the NAO).

We have carried out a preliminary study for October–March. Mean PMSL anomalies in the Atlantic sector tend to be fairly consistent at or near solar maximum, but much less consistent and weak around solar minimum. So we confine our results to high values of TSI. Figure 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; values smaller than 20 % lose significance because of the limited number of winters. Some studies suggest that given phases of the QBO and solar minima may interact to influence North Atlantic winter atmospheric circulation (Anstey and Shepherd, 2014), so this could be a topic for the future.

3.2.5 The Atlantic Multidecadal Oscillation

The AMO is likely to be both largely an 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 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. Figure 11 shows global PMSL and UK and European rainfall anomalies for the winter half year AMO index over the common data availability period 1901–2011 for winter half year AMO values > 1 and < 1 SD calculated for this period, corresponding to warm and cold

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North Atlantic states corrected for the trends in global mean sea surface temperature, respectively (the state in 2014 is warm). The AMO varies mostly interdecadally so any AMO related climate signal is 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 positive and a positive NAO when the AMO is negative. AMO effects on rainfall over much of UK are clearest for the negative AMO 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 implied by Sutton and Dong's results for autumn and spring. So 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 Sect. 2.

Figure 12 comprises boxplots of the various response variables for the winter half year (rainfall, river flow, groundwater and SGI flow, SGI groundwater and SPI), 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 Fig. 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 Sect. 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.

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 Conclusions

There has been a recent considerable improvement in dynamical seasonal weather forecasting models and in our understanding drivers of climate variability in the winter half year. Besides more detailed studies of climate drivers and their intraseasonal impacts, hydrological and groundwater models eventually need to be integrated into seasonal weather forecast models. Because Lowland England is a small area only moderately affected by the winter North Atlantic Oscillation, it may be some time before such integrated models have really useful skill in this region. Accordingly, observationally-based methods taking advantage of the close relationship shown in this paper between atmospheric circulation, rainfall and hydrological anomalies are likely to have a significant role in the medium term.

4.1 General considerations

The predictability of winter droughts in the English Lowlands is a strongly multivariate 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 but it is much more important in summer. 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 Sect. 3. Folland et al. (2012) point out that the magnitude of the drivers we discuss in Sect. 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

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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 Niòas 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). Folland et al. (2009) showed that in summer, the summer NAO is the most prominent atmospheric circulation pattern. 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

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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.

4.2 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. HadISST, 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 reso-

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lution 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). 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 1870, 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.

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 and has shed some new light on the propagation from meteorological through to hydrological drought – in particular lags between the meteorological drought anomalies (SPI) and their response in river flow and groundwater levels. However, this has only been conducted at a broad scale for the English Lowlands – the time lags 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 press). There is a 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 Niño 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 Niña or El 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

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Table 2. Top 20 winter half year La Niòas and English Lowlands rainfall since 1910–1911.

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^{-1}
1988–1989	−1.87	YES	−15.2
1973–1974	−1.82	YES	−9.3
2007–2008	−1.56	NO	1.7
1942–1943	−1.46	NO	2.3
1999–2000	−1.43	NO	−6.8
2010–2011	−1.42	YES	−10.5
1998–1999	−1.39	NO	−15.2
1975–1976	−1.32	YES	−26.0
1970–1971	−1.25	NO	4.2
1916–1917	−1.20	NO	4.5
1949–1950	−1.10	NO	9.3
1984–1985	−1.09	NO	−0.2
1933–1934	−1.05	YES	−19.7
1955–1956	−1.02	NO	−5.7
1924–1925	−0.89	NO	8.7
1938–1939	−0.88	NO	14.7
2011–2012	−0.86	YES	−18.9
1995–1996	−0.85	YES	−10.5
1983–1984	−0.71	NO	1.0
1910–1911	−0.71	NO	8.3

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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.

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 (Supplement 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.

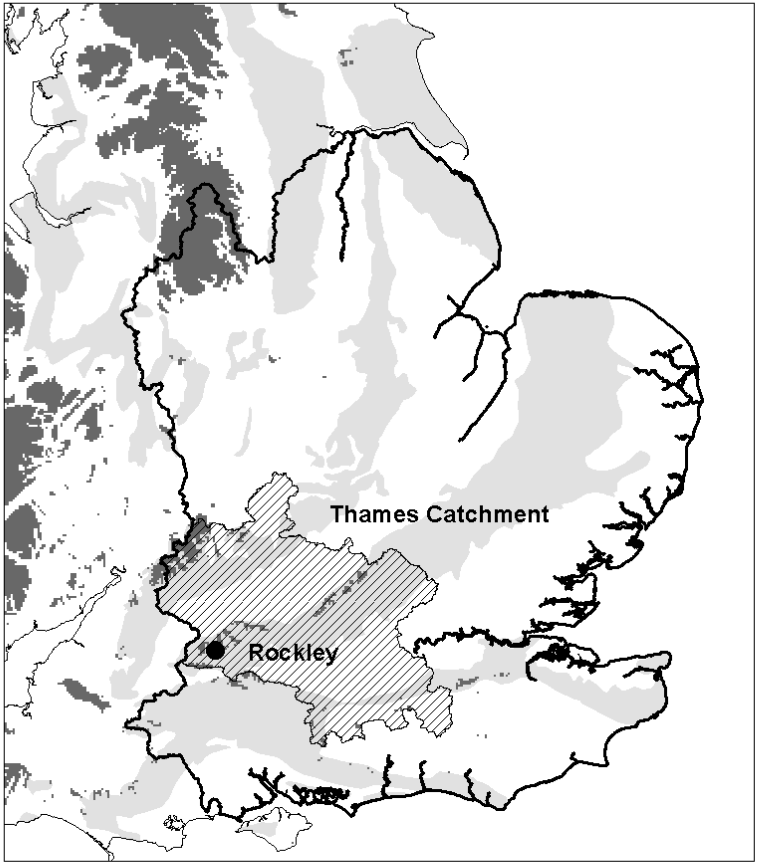

Figure 1. Map of the English Lowlands study region (bold line indicates boundary), also showing the Thames catchment (above the Kingston gauging station) and the location of the Rockley borehole. Map also shows major aquifers (light grey) and upland areas over 200 m (dark grey).

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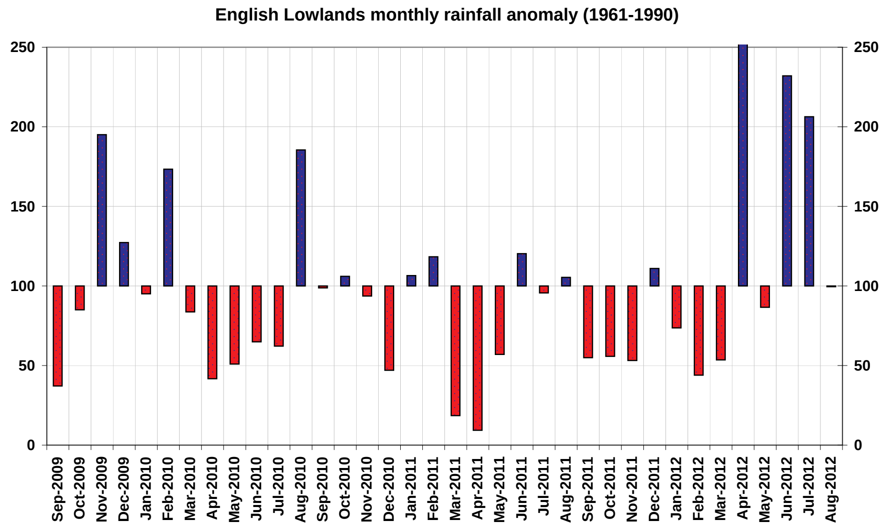

Figure 2. Example of a meteorological drought, April 2010 to March 2012.

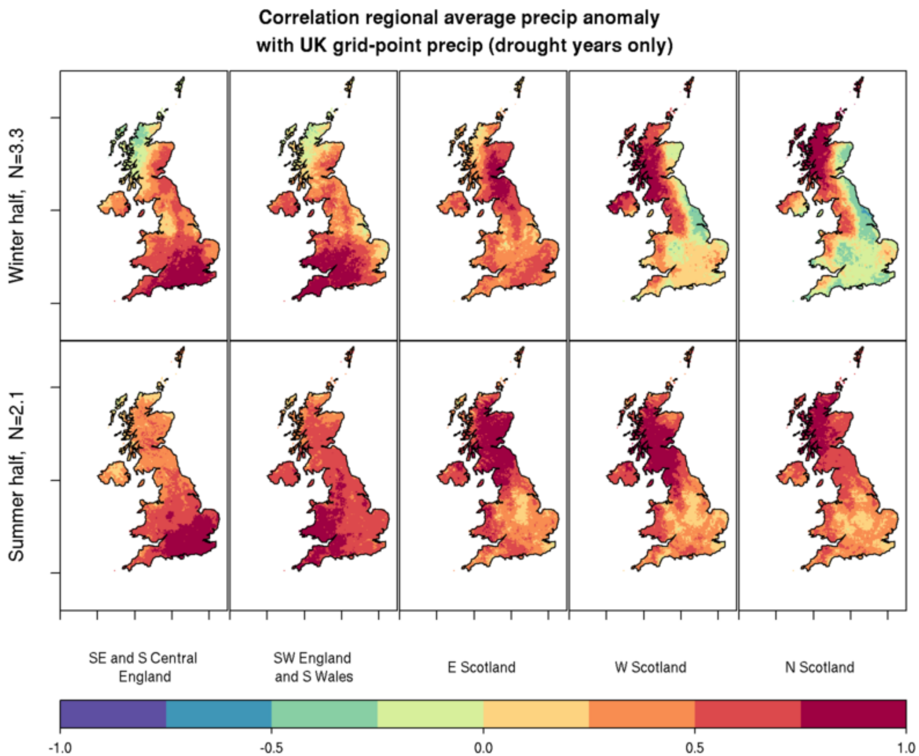

Figure 3. Correlations of designated district average rainfalls with 5 km × 5 km gridded rainfall data elsewhere in UK for winter and summer half years of droughts identified in this paper. *N* is the calculated equivalent number of independent rainfall stations across the UK in Table 1 droughts, a measure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

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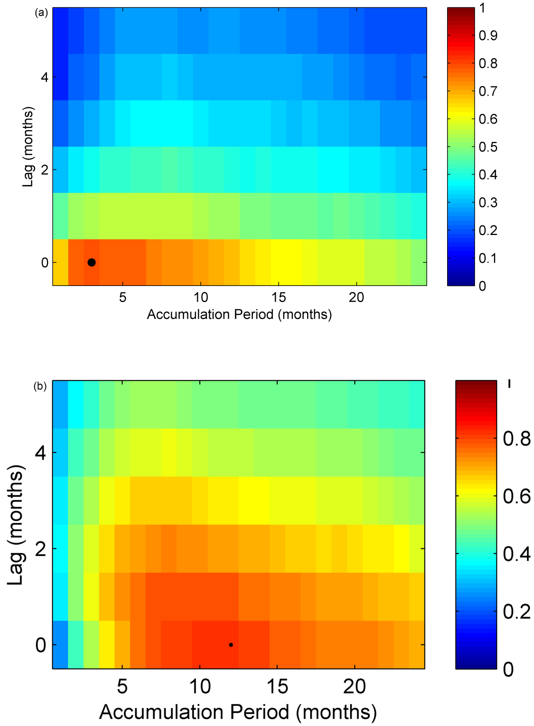

Figure 4. (a) Heatmap of the correlation between lagged English Lowlands river flow SGI over a one month timescale and English Lowlands precipitation as SPI over 1–24 months, with maximum correlation highlighted with black circle. **(b)** Heatmap of the correlation between lagged English Lowlands groundwater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

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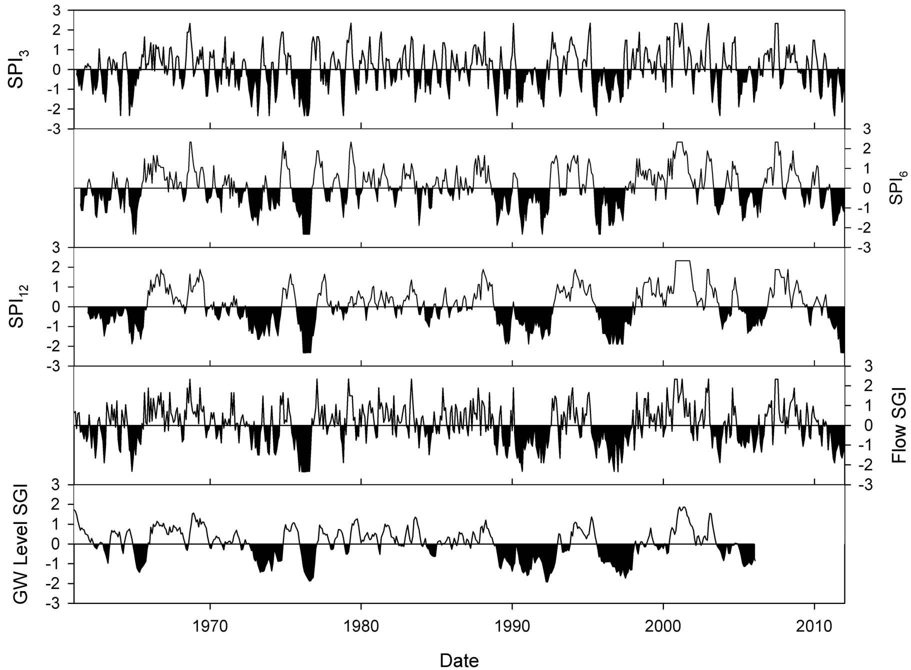

Figure 5. SPI and SGI for regional English Lowlands series, where the first three time series are SPI based on the English Lowlands precipitation time series, with SPI 3 month rainfall accumulation, SPI 6 month rainfall accumulation and SPI 12 month rainfall accumulation; the latter two are SGI for the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

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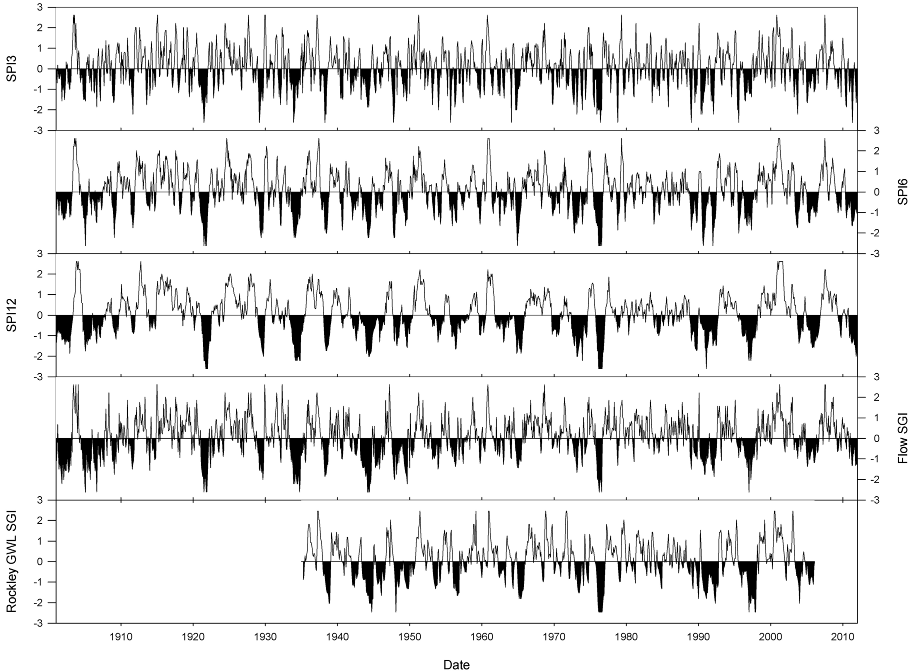

Figure 6. Figure 6 shows the SPI and SGI series for the long Thames record, the long Thames catchment rainfall series and the Rockley borehole.

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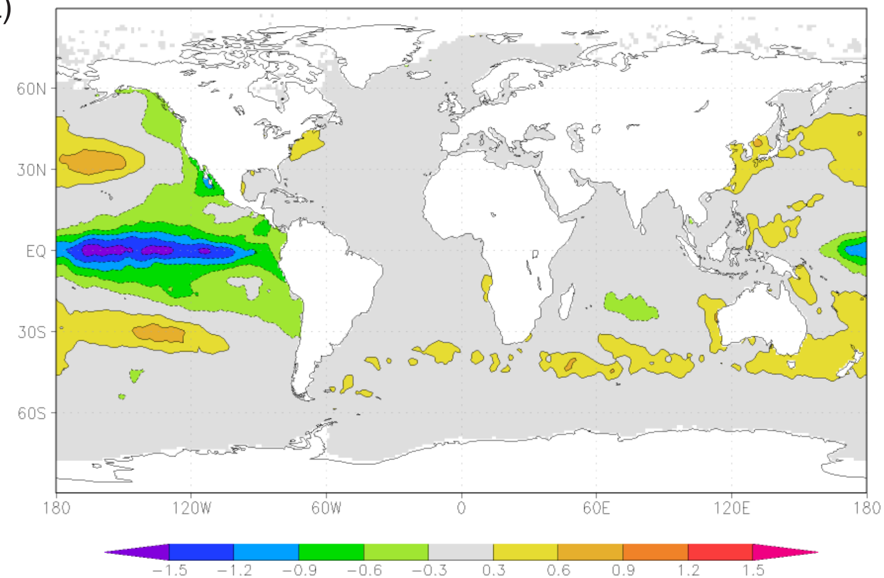
(a)



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(b)

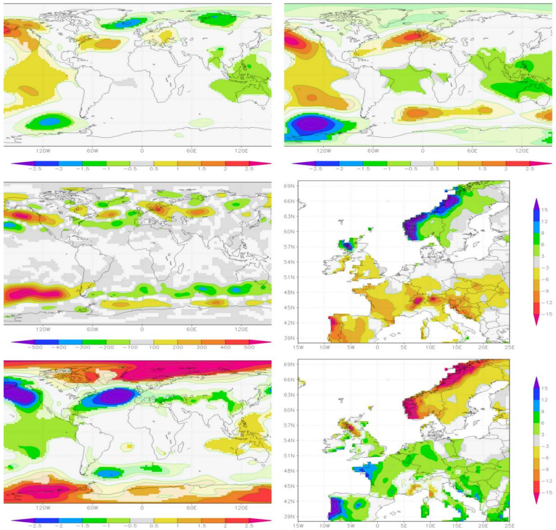

Figure 7. **(a)** Composite global SST anomalies from 1961–1990, winter half year, over 1901–2013 when Nino 3.4 anomalies < −1.0 °C. **(b)** Top panels: global PMSL anomalies (hPa) from the 20th century reanalysis averaged over winter half year for La Niñas measured by SST < −1 SD over Nino 3.4, corresponding to a 1961–1990 SST anomaly < −0.92 °C, for two independent epochs 1876–1950 (left) and 1951–2009 (right). The SD is for 1951–2010. Central panels (left): global storminess anomalies, 1951–2013 measured by anomalies of 2–7 day band pass variance of 500 hPa height (dm^2), (right) west European rainfall anomalies (mm $month^{-1}$) 1901–2011 for La Niñas for winter half year. Bottom panels (left): as top right panel for moderate El Niños (anomalies of 0.92 °C < Nino 3.4 < 1.5 °C) (right) as central right panel but for moderate El Niòos. Dark colours are locally significant at the 5 % level. Rainfall from the Mitchell and Jones (2005) 0.5° × 0.5° degree data set, as it is for Figs. 9–12.

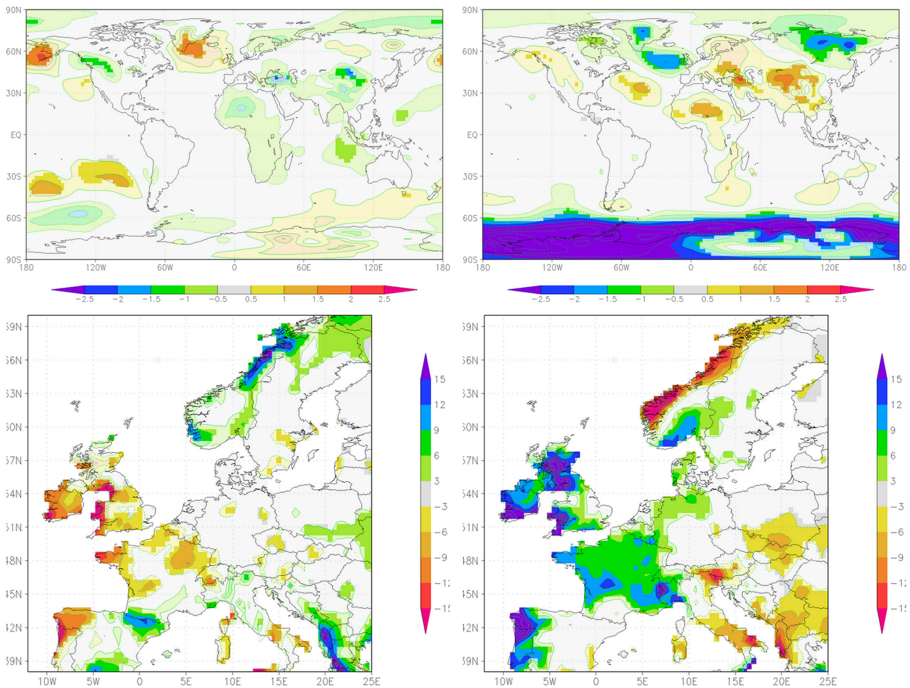

Figure 8. (top left) Global PMSL anomalies (hPa) in winter half year for a tripole SST index < −1 SD; (top right) > 1 SD in the previous may. (bottom left) Rainfall anomalies in winter half year (mm month^{-1}) over UK and nearby Europe for tripole SST index < −1 SD. (Bottom right) for > 1 SD. Areas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

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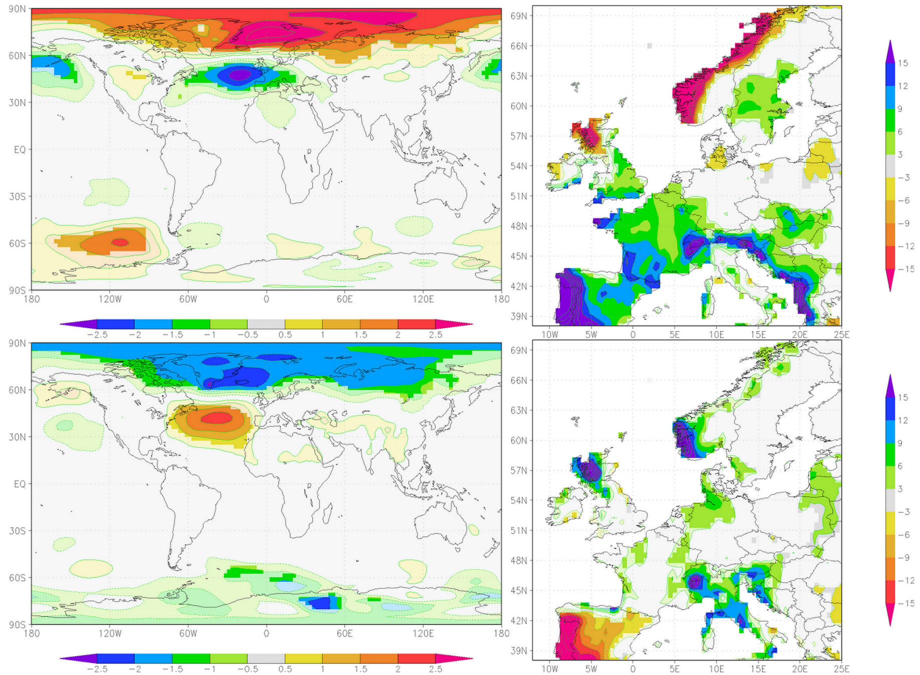

Figure 9. (top left) Near global PMSL anomalies (hPa) in winter half year for most easterly QBO 15 % of 30 hPa equatorial stratospheric winds (1953–1954 to 2012–2013). (top right) Rainfall anomalies for the top 15 % most easterly of all equatorial winds. (bottom left) As top left but for the 15 % most 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.

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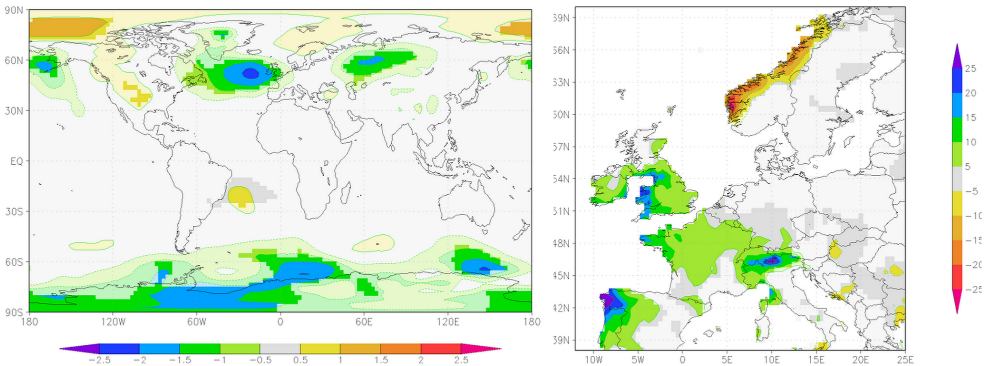

Figure 10. (left) Near global PMSL anomalies (hPa) in winter half year for TSI values in the highest 20 % of the its winter half year distribution over 1948–2011. Earlier years not used as solar cycle mostly varied at an averaged reduced level of total solar radiation. (right) rainfall anomalies (mm month^{-1}) over UK and nearby Europe. Areas significant at the 5 % level are darker coloured.

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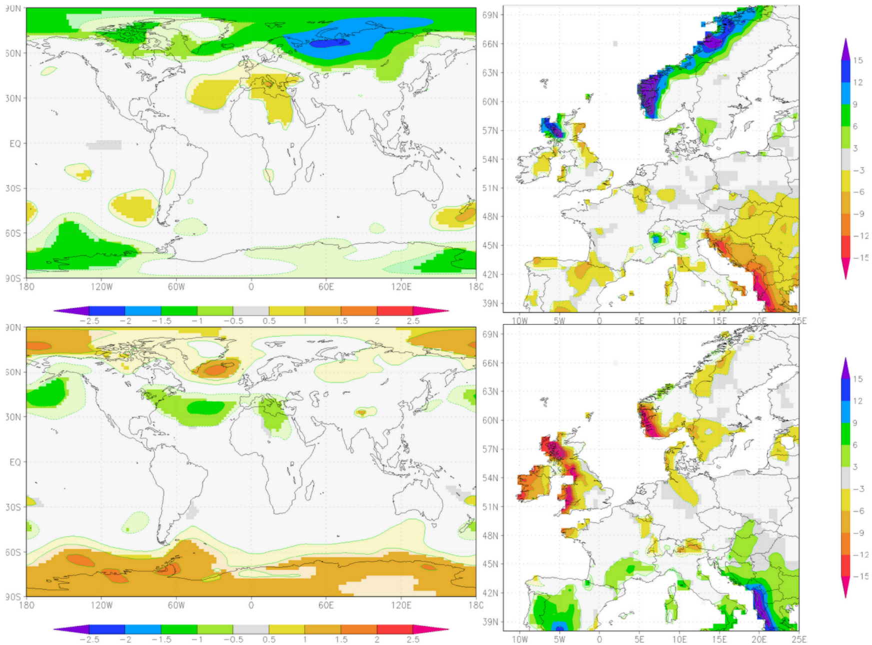
(a)



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(b)

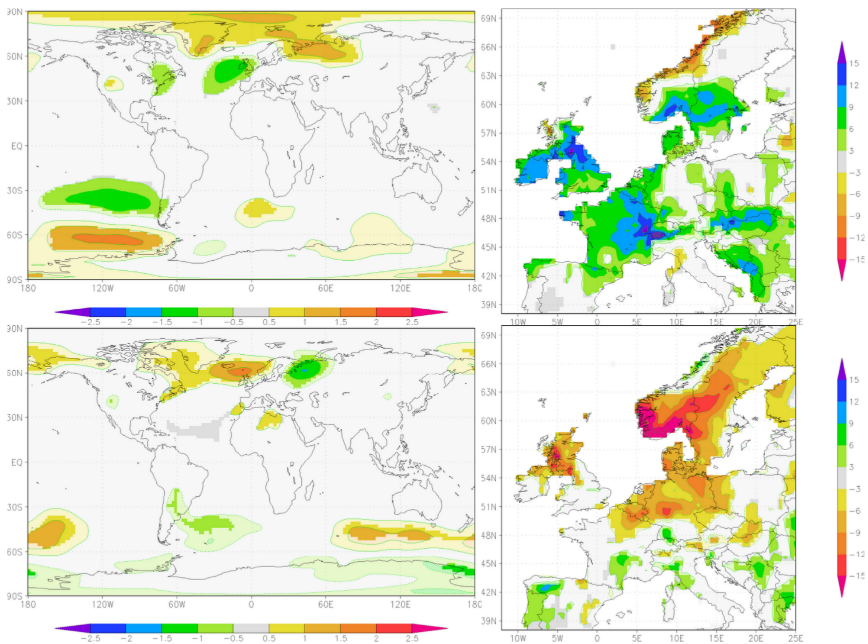

Figure 11. (a) (top left) Near global PMSL anomalies (hPa) in winter half year for monthly AMO index values < −1 SD calculated over 1871–2013. (top right) rainfall anomalies (mm month^{-1}) for AMO index values < −1 SD. (bottom left) Near global PMSL anomalies for AMO index values > 1 SD (bottom right) Rainfall anomalies (mm month^{-1}) for AMO Index values > 1 SD. Areas significant at the 5 % level are darker coloured. PMSL is from the 20CR. **(b)** (top left) Near global PMSL anomalies (hPa) in summer for monthly AMO index values < −1 SD calculated over 1871–2013. (top right) Rainfall anomalies (mm month^{-1}) for AMO index values < −1 SD. (bottom left) Near global PMSL anomalies for AMO index values > 1 SD (bottom right) rainfall anomalies (mm month^{-1}) for AMO Index values > 1 SD. Areas significant at the 5 % level are darker coloured. PMSL is from the 20CR.

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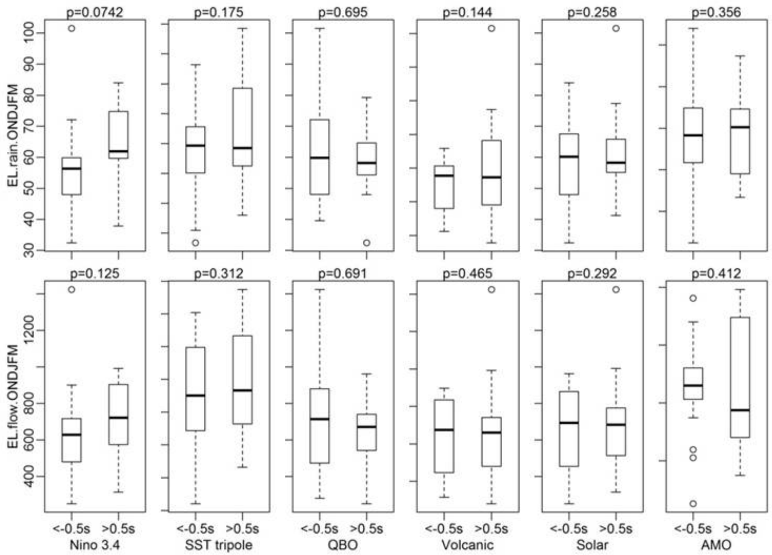
(a)



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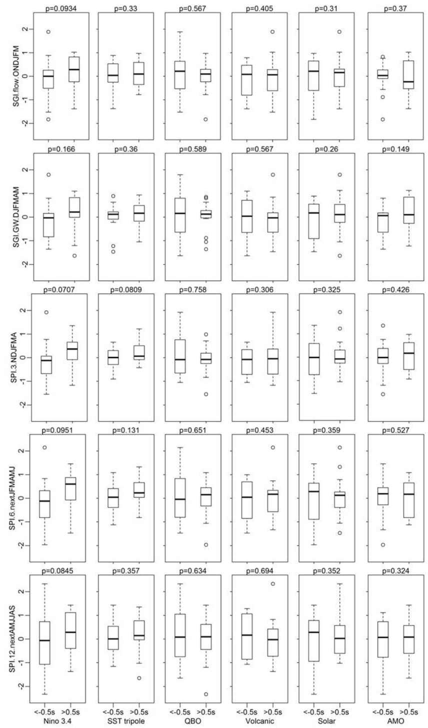

Figure 12. (a) Box plots of English Lowland response variables for the October to March winter half year (English Lowlands areal rainfall and total flow), for low (< −0.5 SD) and high (> 0.5 SD) values of different drivers (Niño 3.4, IPO, TSI, May SST tripole, AMO, stratospheric aerosol loadings and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