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Global meteorological influences on the record UK rainfall of winter 2013–14

Jeff R Knight¹, Anna Maidens¹, Peter A G Watson², Martin Andrews¹, Stephen Belcher¹, Gilbert Brunet^{1,3}, David Fereday¹, Chris K Folland^{1,4,5,6}, Adam A Scaife¹ and Julia Slingo¹¹ Met Office, FitzRoy Road, Exeter EX1 3PB, United Kingdom² Atmospheric, Oceanic and Planetary Physics, University of Oxford, Oxford, United Kingdom³ Now Environment and Climate Change Canada, Trans-Canada Highway, Dorval, Québec H9P 1J3, Canada⁴ School of Environmental Sciences, University of East Anglia, Norwich, United Kingdom⁵ Department of Earth Sciences, University of Gothenburg, Gothenburg, Sweden⁶ International Centre for Applied Climate Sciences, University of Southern Queensland, Toowoomba, Queensland, AustraliaE-mail: jeff.knight@metoffice.gov.uk**Keywords:** winter 2013–14, floods, UK rainfall, atmospheric circulation, Rossby waves, climate changeSupplementary material for this article is available [online](#)

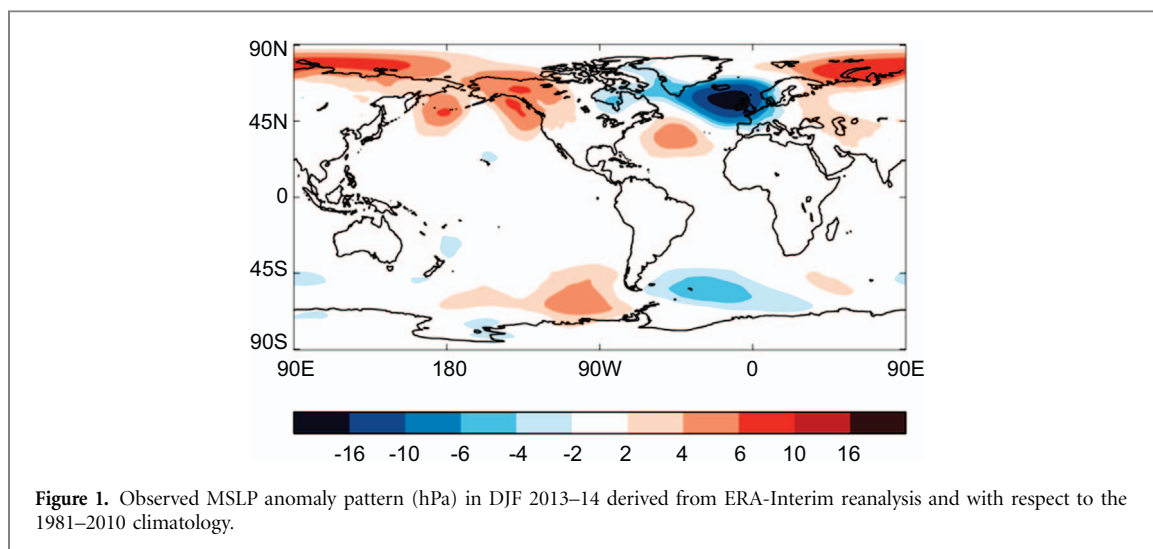
Abstract

The UK experienced record average rainfall in winter 2013–14, leading to widespread and prolonged flooding. The immediate cause of this exceptional rainfall was a very strong and persistent cyclonic atmospheric circulation over the North East Atlantic Ocean. This was related to a very strong North Atlantic jet stream which resulted in numerous damaging wind storms. These exceptional meteorological conditions have led to renewed questions about whether anthropogenic climate change is noticeably influencing extreme weather. The regional weather pattern responsible for the extreme UK winter coincided with highly anomalous conditions across the globe. We assess the contributions from various possible remote forcing regions using sets of ocean–atmosphere model relaxation experiments, where winds and temperatures are constrained to be similar to those observed in winter 2013–14 within specified atmospheric domains. We find that influences from the tropics were likely to have played a significant role in the development of the unusual extra-tropical circulation, including a role for the tropical Atlantic sector. Additionally, a stronger and more stable stratospheric polar vortex, likely associated with a strong westerly phase of the stratospheric Quasi-Biennial Oscillation (QBO), appears to have contributed to the extreme conditions. While intrinsic climatic variability clearly has the largest effect on the generation of extremes, results from an analysis which segregates circulation-related and residual rainfall variability suggest that emerging climate change signals made a secondary contribution to extreme rainfall in winter 2013–14.

1. Introduction

Between December 2013 and March 2014 many areas of the United Kingdom (UK) experienced severe river and groundwater flooding (Huntingford *et al* 2014, Muchan *et al* 2015). In the low-lying Somerset Moors, flooding to a depth of several metres persisted over an area of up to 65 km² throughout much of this period. Widespread and persistent impacts gave the impression that events in this winter were unprecedented. While the occurrence of flooding depends on a range of factors, including flood management, the rainfall

itself was consistent with this view. UK average precipitation for the meteorological winter (December to February or DJF) was 545 mm (Kendon and McCarthy 2015), which is the wettest winter in the series starting in 1910 and more than 50 mm wetter than the previous record (485 mm in 1994–5). Average values for England, Wales and Scotland were also individual records in their respective series. Looking further back, 2013–14 is the wettest winter (456 mm) in the England and Wales Precipitation (EWP) series (Alexander and Jones 2001), which starts in 1766 (albeit based on fewer rainfall stations than the various



series from 1910). The extreme UK rainfall was part of a pattern of high precipitation across much of the European Atlantic seaboard including Portugal, France, and Ireland, as well as in Southern Scandinavia, Northern Italy and the Alps. Flooding was additionally experienced in a number of these places.

The immediate cause of the record UK rainfall was a prolonged sequence of winter cyclones tracking over or close to the British Isles from mid-December onwards (Christidis and Stott 2015). The mean storm track in winter 2013–14 was therefore displaced southwards relative to its climatological position. Additionally, many of these systems were very intense, resulting in the occurrence of damaging winds alongside heavy rain. The seasonal average mean sea level pressure (MSLP) shows highly negative anomalies (figure 1) over much of the North West Atlantic region, reaching -19.6 hPa or -4.1 standard deviations to the west of the British Isles. MSLP anomalies across a broad region of the North East Atlantic were the most extreme for over 100 yr (see supplementary information available at stacks.iop.org/ERL/12/074001/mmedia). Anomalously high MSLP values occurred southeast of the region of low MSLP, and the pattern projects onto the positive winter North Atlantic Oscillation (NAO; Hurrell 1995), the principal mode of year-to-year variability in winter atmospheric circulation in the North Atlantic sector. The exact pattern is displaced southwards in the Eastern Atlantic compared to an NAO pattern derived using data from many winters, and projects onto the East Atlantic pattern (Moore and Renfrew 2012). The NAO index (the anomalous difference between Iceland and Azores station MSLP) for winter 2013–14 is still $+11.3$ hPa or $+1.3$ standard deviations, however (for a 1981–2010 climatology period). The winter NAO has been linked to a range of external influences, such as the state of the El Niño–Southern Oscillation (Fereday *et al* 2008) and the stratosphere (Kidston *et al* 2015), and it seems probable that the extreme North Atlantic circulation observed in winter 2013–14 was made more

likely by remote factors present in the climate system at the time. Any explanation for the extreme UK rainfall based on these remote drivers must both account for the positive state of the NAO and the southward displacement of the storm track over the eastern North Atlantic Ocean.

The extreme atmospheric circulation in the North Atlantic region coincided with other notable seasonal anomalies across the globe (see online supplementary information), including what was likely the coldest winter over parts of North America in a century and unusually warm marine conditions and high rainfall in the western tropical Pacific Ocean. It has been suggested that anomalies in the tropical West Pacific could have given rise to the observed conditions over North America, and thereby influenced the North Atlantic sector (Slingo 2014, Palmer 2014, Huntingford *et al* 2014). Numerical climate model experiments have been performed to test these links (Hartmann 2015, Lee *et al* 2015), and show that North Pacific and North American anomalies could have been stimulated by the higher-than-usual tropical Pacific SSTs. Watson *et al* (2016) also demonstrate the potential for such links (albeit weaker than observed) in models forced with observed SSTs. Their analysis also demonstrated, however, that the atmospheric dynamical response to imposed SSTs over the tropical and sub-tropical West Pacific was not realistic compared to reanalysis data. Dynamical analysis of observations nevertheless shows the potential for the region to produce strong extra-tropical teleconnections during January. In addition, the stratospheric polar vortex was strong in winter 2013–14, with an anomalously westerly polar night jet (see supplementary information). A strong polar vortex is often associated with a positive NAO (Baldwin and Dunkerton 2000), as was observed. The strong westerly phase of the QBO prevalent at the time likely contributed to the polar vortex strength; the QBO has long been proposed as a driver of winter NAO variability (Ebdon *et al* 1975) and more recent observational analyses confirm earlier indications

(Scaife *et al* 2014). Lastly, Kug *et al* (2015) have suggested that near-surface warmth in parts of the Arctic, presumably related to the observed decline in Arctic sea-ice, could create an atmospheric circulation response over North America and the North Atlantic Ocean. They show that this warmth was particularly marked in winter 2013–14.

Although UK rainfall in winter 2013–14 was particularly remarkable in exceeding previous records by so great a degree, there have been a number of extremely wet UK months or seasons in recent years. The current UK average precipitation (Perry and Hollis 2005) records for six calendar months and both the autumn and winter seasons have been set since the year 2000. In addition, six of the seven wettest calendar years have occurred in this time. Many of these periods are associated with episodes of widespread flooding. In this context, the even more extreme rainfall in winter 2013–14 clearly prompts the question of whether these impacts are related to changes in the global climate and in particular anthropogenic climate change. Indeed, the floods in autumn 2000 have been attributed, in part, to increasing greenhouse gases (Kay *et al* 2011, Pall *et al* 2011). Climate warming can produce more extreme rainfall either via an increased capacity for atmospheric moisture (the ‘thermodynamic mechanism’) or by changes in atmospheric circulation patterns (the ‘dynamic mechanism’—Shepherd *et al* 2014). There is potentially a greater effect from the latter mechanism as atmospheric circulation anomalies are the primary drivers of rainfall anomalies in mid-latitudes. Patterns of circulation are dynamically interlinked at a near-global scale, however, meaning that any single location is potentially influenced by anomalies across the globe. This has made understanding and modelling the response of global circulation patterns to future levels of greenhouse gas forcing a particular challenge (Fereday *et al* 2017). Nevertheless, Schaller *et al* (2016) use climate model simulations to suggest that both dynamic and thermodynamic climate change increased UK rainfall in winter 2013–14, although this finding may be dependent on the model used. In contrast, Rasmijn *et al* (2016) show that a future winter with a similar sub-polar jet position might be expected to produce less excessive rainfall and storminess due to weaker baroclinic gradients.

In this paper we relate the deep negative North East Atlantic MSLP anomaly in winter 2013–14 to remote phenomena that were present in the climate system at the time. This is done in part by looking at potential drivers of the positive NAO state observed in this winter, but also by seeking influences giving the particular southward-shifted pattern that was observed. By performing coupled ocean–atmosphere model experiments, we attempt to identify the likely influences that made the winter so extreme in terms of circulation and hence precipitation. Lastly, we attempt to infer the contributions to the extreme rainfall from

the thermodynamic mechanism and atmospheric circulation, which is useful in giving an indication of the likelihood of such extremes in future.

2. Constrained simulations of winter 2013–14

In this section, we use ensembles of seasonal-length numerical ocean–atmosphere model experiments to test to what degree observed conditions in various parts of the atmosphere affected the Northern Hemisphere extra-tropical circulation. To do this, we use a relaxation technique, which constrains the simulated state to be similar to the observed state within a specified spatial domain while leaving remaining parts of the simulation to evolve as directed by the model’s dynamics and physics. With this methodology, model fields are modified at each time step of the simulation by adding a fraction of the difference between observed and simulated values everywhere within the chosen domain. This fraction determines the effective timescale for the relaxation of model fields towards observations. The relaxation technique has been successfully applied for instance in performing case studies of particular years (Jung *et al* 2008, Greatbatch *et al* 2015, Watson *et al* 2016) and in investigating factors driving climate variability (Gollan *et al* 2015).

Relaxation experiments were performed with the Met Office coupled atmosphere–ocean model, HadGEM3 (Hewitt *et al* 2011) and the European Centre for Medium Range Forecasts (ECMWF) model (Watson *et al* 2016). Both models have an effective atmospheric resolution of between 50 and 100 km. The Met Office model has an oceanic resolution of 0.25 degrees while the ECMWF model has a resolution of 1 degree. Five experimental configurations were employed with HadGEM3, relaxing the whole tropical tropospheric band, the tropical Atlantic (15°–75°W), tropical West Pacific (120°–170°E), the global stratosphere and the whole tropics plus the global stratosphere. For the tropical configurations, relaxation was performed over the latitude range 22.5°S to 22.5°N, and the boundary of the tropospheric and stratospheric domains is at 20 km altitude. The temperature T and horizontal winds u , and v are relaxed towards analysed values with an e-folding timescale of 6 h. The relaxation timescale is tapered to zero over 12.5° of longitude at the edges of the tropical Atlantic and tropical West Pacific domains to minimise any ‘shock’ when the predominantly zonal-moving trade winds impinge on the domains. For the ECMWF model, a tropical relaxation experiment and a tropical Atlantic (90°W–20°E) experiment were performed with a latitude range of 20°S to 20°N. Relaxation is performed as described in Watson *et al* (2016).

Within the five HadGEM3 experimental configurations, two 70-member ensembles of simulations

were performed. The first uses relaxation towards the Met Office's real-time numerical weather prediction analysis covering the period November 2013 to February 2014, while the second is a control ensemble using ERA-Interim data (Dee *et al* 2011) for equivalent periods in the 14 winters 1996–7 to 2009–10. For the purpose of this study, we treat the two sources of relaxation data as equally accurate representations of the atmospheric state, constrained as they are by extensive *in situ* and satellite observations. Using large ensemble sizes ensures that the effect of chaotic variability in the extra-tropical atmosphere is averaged out, such that all the simulated features we show here are statistically significant to a high level of confidence. We use a control ensemble to remove any systematic effect of using relaxation unrelated to the specific conditions in winter 2013–14. The difference between ensemble means of the 2013–14 and control ensembles is interpreted as the impact of imposing realistic conditions within a specific relaxation domain. The ECMWF 2013–14 ensembles have 28 members while control ensembles with 64 members were additionally performed. Relaxation is towards ERA-Interim in both.

The HadGEM3 ensembles were initialised using realistic states (MacLachlan *et al* 2014) from ERA-Interim and output from the Met Office Forecast Ocean Assimilation Model (FOAM). Initial conditions for 1st November of the range of years 1996–2009 are used in both ensembles, meaning that any signals related to the initial conditions are, to a first approximation, cancelled by averaging over the ensemble. The set of initial states are cycled through the ensemble, and members with the same initial conditions are given different values of an initial seed value of the model's stochastic physics scheme, which adds different random perturbations during the simulation, ensuring these members diverge. This experimental design provides signals that are, to a good approximation, only a result of the observationally-based data introduced by relaxation. The ECMWF ensembles, in contrast, are initialised with conditions for 1st November before the winter of the relaxation data applied in each case. In this way, these ensembles are constrained seasonal re-forecasts. This is not expected to create difficulty in comparing with the HadGEM3 results since the ensemble means of mid-latitude North Atlantic winter mean anomalies in initialised seasonal forecasts tend to be considerably weaker than those we obtain in the relaxation experiments performed here.

The results of the relaxation experiments are shown in figure 2. We first consider ensembles with relaxation of the global stratosphere (upper left). The difference between the DJF average ensemble mean MSLP fields shows moderate negative anomalies (2–4 hPa) across the whole Arctic, reminiscent of the pattern of the Arctic Oscillation (AO; Thompson and Wallace, 1998). Anomalies are slightly stronger in the Atlantic sector with positive anomalies to the south, reminiscent of the NAO. Differences elsewhere

are small. The overall NAO signal is 5–6 hPa; about half the NAO anomaly observed in winter 2013–14. The circulation response is consistent with the strong polar vortex in winter 2013–14 and is expected given previous studies showing the link between vortex strength and tropospheric circulation (Kidston and Chiba 1995, Baldwin and Dunkerton 2001, Scaife and Knight 2008).

The effect of relaxing the tropical West Pacific state in the model to winter 2013–14 observations (upper right in figure 2) is to produce cyclonic anomalies in the North Pacific and a dipole anomaly in the North Atlantic which projects onto the negative NAO. As can be seen in the seasonal average anomaly derived from reanalysis (bottom right), however, the simulated features are spurious. This lack of agreement is consistent with the results of Watson *et al* (2016) who also show limited success in simulating the North Atlantic anomaly by relaxation of the West Pacific. We widen our search, therefore, by examining the response to relaxation of the whole tropical band (second row in figure 2). In contrast, this reveals features with a marked resemblance to the observed anomalies. These include positive MSLP differences in the North Pacific and a dipole anomaly in the North Atlantic. Similar features are simulated in each model. In HadGEM3, a deep (>10 hPa) centre of low MSLP is produced in the northern North Atlantic, although it does not extend far enough across Europe. The ECMWF model better simulates the position of the low pressure but not the magnitude. Overall, the global pattern correlations between the models and observations are +0.51 (HadGEM3) and +0.62 (ECMWF), but this rises to +0.79 and +0.77 respectively—a high level of similarity—in the Atlantic sector. The agreement of both models with observations strongly suggests that tropical forcing was key to generating the observed North Atlantic anomalies.

We next ask which parts of the tropics are responsible for the extra-tropical signals by examining the tropical Atlantic (third row in figure 2). Many of the features highlighted by pan-tropical relaxation are seen in the tropical Atlantic ensembles, principally high MSLP in the North Pacific (albeit weakly in the ECMWF model) and a marked cyclonic feature in the North Atlantic. Although the North Atlantic pattern is somewhat weaker than for the all tropics case in HadGEM3, the results suggest that the tropical Atlantic contributes significantly to the overall response from tropical forcing. We have also suggested that the stratosphere played a role in shaping the winter of 2013–14, so we perform ensembles to estimate the response to constraining the tropics and the stratosphere together. The MSLP difference patterns obtained (lower left panel in figure 2) look similar to those from the tropics only comparison and to the observed anomalies. While the North Atlantic cyclonic anomaly is less intense than in the tropics-only results, it extends further east and there is a

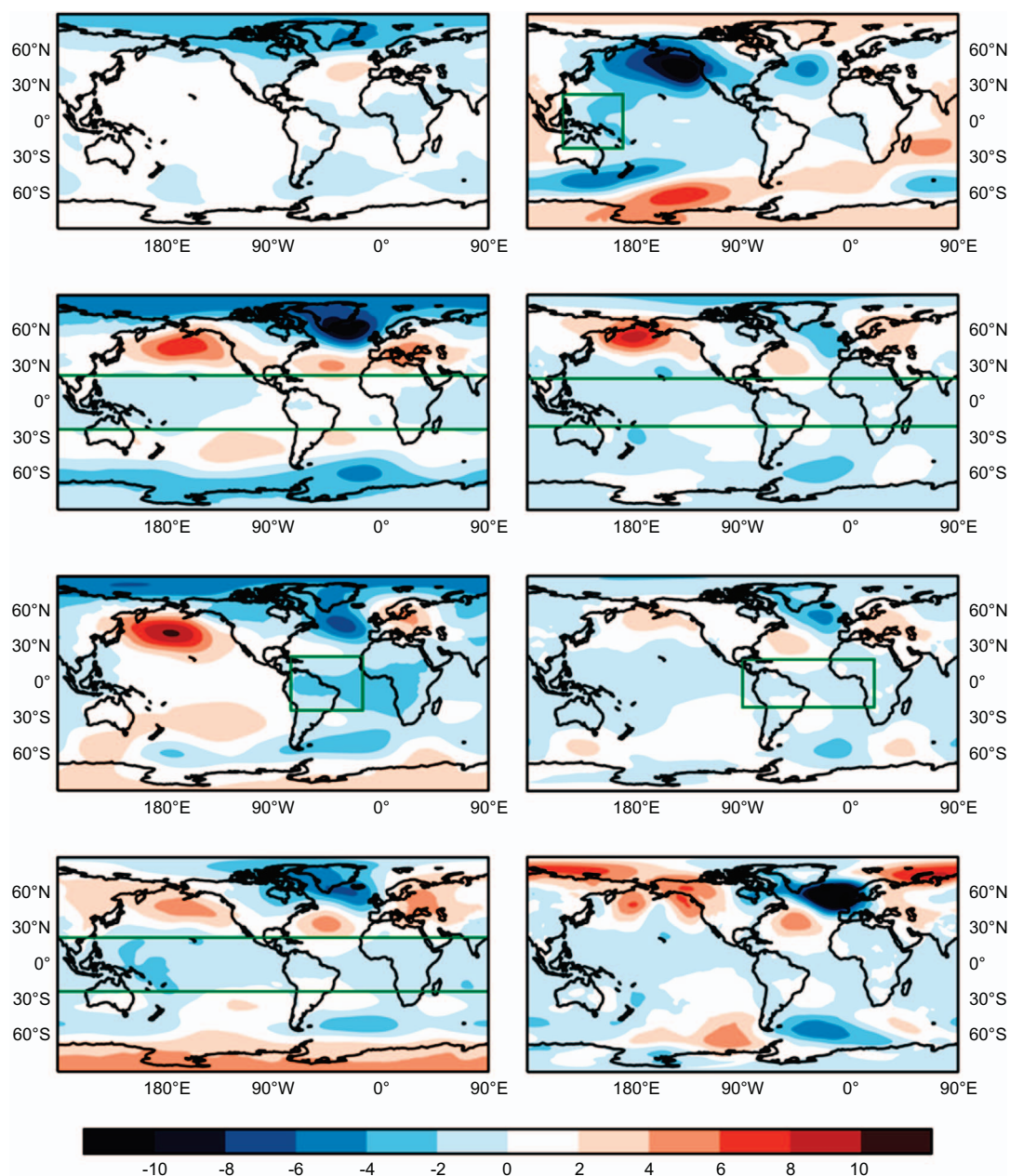


Figure 2. Simulated influences on European winter 2013–14. DJF MSLP differences (hPa) between the ensemble means of pairs of ensembles for the various relaxation domains listed as follows. Upper left: HadGEM3 global stratosphere (above 20 km); upper right: HadGEM3 tropical West Pacific (120°–170°E); second row, left: HadGEM3 whole tropics (180°W–180°E); second row, right: ECMWF whole tropics; third row, left: HadGEM3 tropical Atlantic (15°–75°W); third row, right: ECMWF tropical Atlantic (90°W–20°E); bottom left: whole tropics plus global stratosphere. The latitude limits for each of the tropical domains is 22.5°S to 22.5°N for HadGEM3 and 20°N to 20°S for ECMWF and these are outlined in green. The 2013–14 DJF MSLP anomaly from ERA-Interim is shown in the bottom right panel.

greater east to west gradient over the North Atlantic, which is one of the key features that is needed to explain the extreme conditions in Western Europe. In addition, there is an improved pressure tendency over the Arctic. Combining the two influences produces the best match to the observed pattern of anomalies with pattern correlations of +0.60 globally and +0.81 in the Atlantic sector.

Insights into the dynamics in the simulations of the winter can be obtained by examining upper tropospheric responses. We compute the difference between the 200 hPa streamfunction in the simulations relaxed to 2013–14 and the control ensemble

(figure 3), removing the zonal mean differences to highlight zonally asymmetric patterns of response. For the stratospheric case (upper left), there is very little effect since changes in the highly zonal stratospheric flow predominantly produce a zonal tropospheric response. The relaxation of the tropical West Pacific does introduce spatial changes to the upper flow, but as for the surface response, the patterns are unlike those in the reanalysis (bottom right). Relaxation of the whole tropics (second row) produces upper tropospheric patterns in both models that are very similar to those observed (pattern correlations of +0.78 in HadGEM3 and +0.86 in the ECMWF model

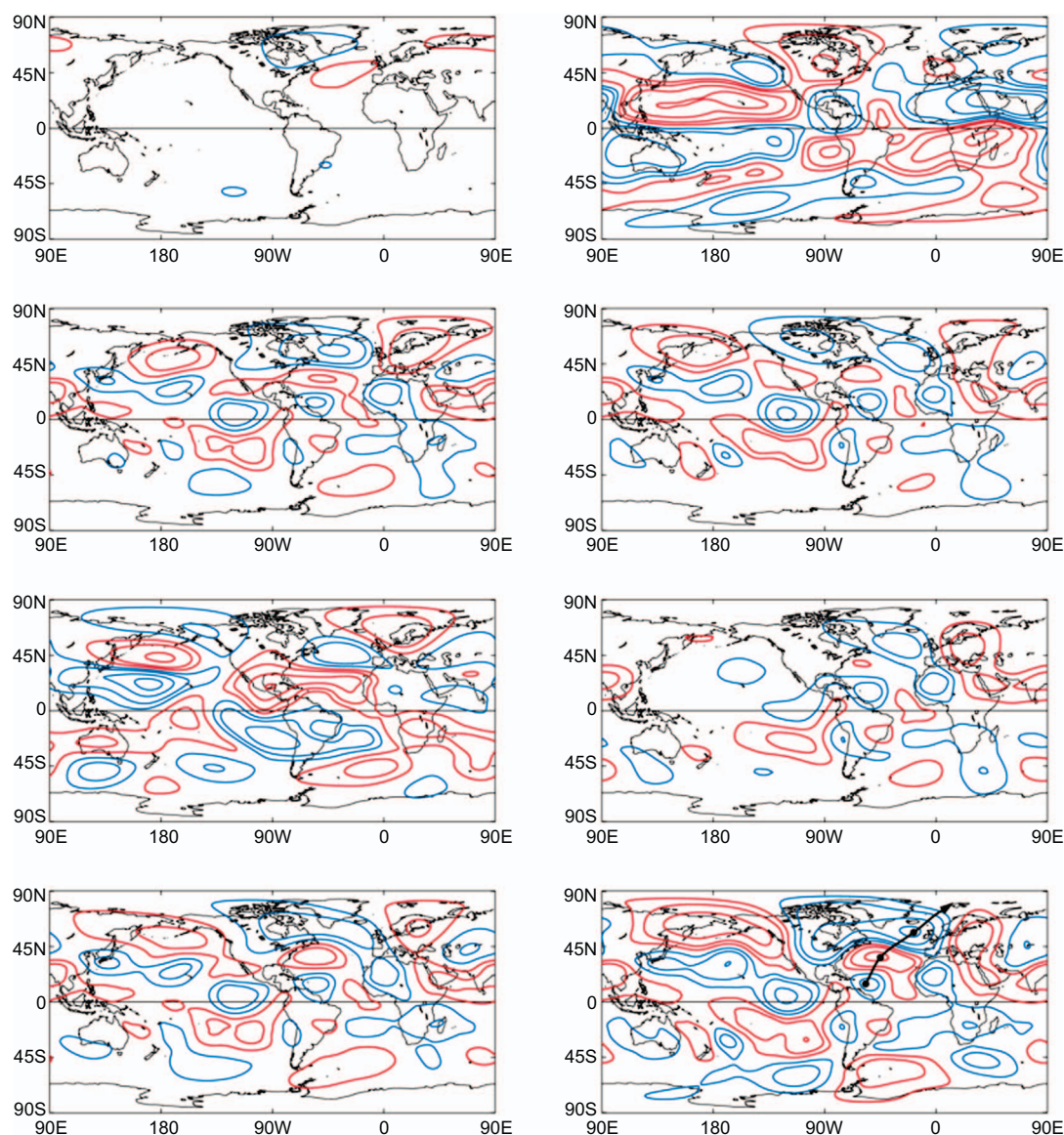


Figure 3. As figure 2, but for the azonal component of the DJF mean 200 hPa streamfunction. Contours plotted are $\pm[2, 5, 10, 15] \times 10^6 \text{ m}^2 \text{ s}^{-1}$. Cyclonic (negative) anomalies are plotted with blue contours and anticyclonic (positive) anomalies are plotted with red contours. For the panel showing the reanalysis (bottom right), the propagation path of the Rossby wave with a zonal wave number 2 obtained from a ray tracing calculation is shown by the curved black arrow. The ray is initiated from the first point on the arrow, which is marked by a black dot, and is close to the centre of the streamfunction anomaly west of the Caribbean Sea. Points along the ray closest to additional streamfunction maxima and minima are also shown with dots.

globally, with similar values in the Atlantic sector). It is straightforward to match almost all of the centres in the streamfunction response to those in the reanalysed winter mean. In particular, the alternating sequence of upper tropospheric cyclonic and anticyclonic features over the North Atlantic Ocean suggests that a Rossby wave emanating from the tropics contributed to the mid-latitude response. This is confirmed by the response to relaxation over the tropical Atlantic alone. Here, wave trains are found emanating from the region into the northern hemisphere. The results from the sets of ensembles to show the combined effect of tropical and stratospheric relaxation are similar (pattern correlations of +0.85 globally and +0.83 in the Atlantic sector) to those from tropical relaxation alone. The existence of a Rossby wave from the tropical Atlantic is further supported by a ray tracing

calculation (Scaife *et al* 2016) which shows that waves with zonal wave number 2 initiated at the edge of the Atlantic tropics would propagate directly through the centres of action of the observed North Atlantic streamfunction anomalies (bottom right in figure 3).

3. Dynamical analysis of tropical forcing

The possible mechanism by which the tropical Atlantic sector contributed to the extratropical circulation anomalies in winter 2013–14 is examined using observational and reanalysis data. In figure 4 we show 2013–14 precipitation anomalies from the Global Precipitation Climatology Project (GPCP) v2.2 dataset (Adler *et al* 2003) compared to the 1981–2010 climatology. Upper tropospheric heating by tropical convection is potentially important for forcing Rossby

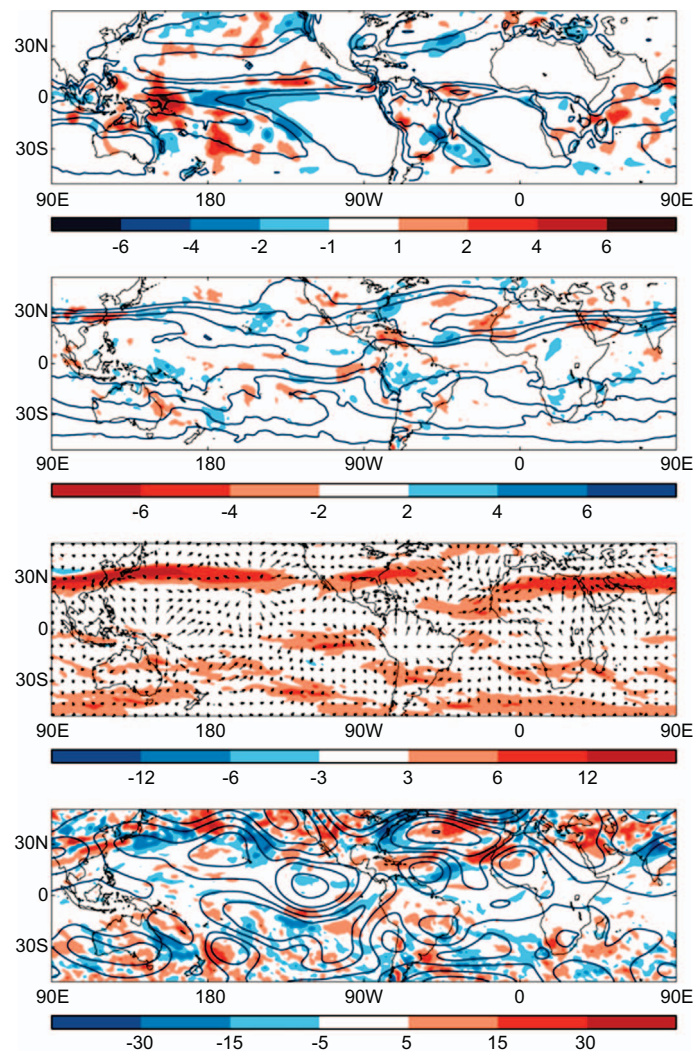


Figure 4. Dynamical analysis of winter 2013–14 using observations and reanalysis. Top panel: DJF precipitation anomaly using the GPCP dataset and a climatological period of 1981–2010. The precipitation climatology is shown using contours of 2, 4 and 8 mm day⁻¹. Second panel: winter mean 200 hPa divergence anomaly (colours, units of 10^{-6} s^{-1}) and absolute vorticity (contours at $\pm[2, 4, 7, 10] \times 10^{-11} \text{ s}^{-1}$). Third panel: meridional gradient of the winter mean 200 hPa absolute vorticity (colours, units of $10^{-11} \text{ s}^{-1} \text{ m}^{-1}$) and anomalous divergent wind (vectors). Bottom panel: 200 hPa winter mean anomalous Rossby wave source (colours, units of 10^{-11} s^{-2}) and zonal component of the anomalous streamfunction (contours of $\pm[2, 5, 10, 15] \times 10^6 \text{ m}^2 \text{ s}^{-1}$). Positive values of the Rossby wave source correspond to a cyclonic tendency and vice versa.

wave propagation into the extra-tropics (Sardeshmukh and Hoskins 1988). The large precipitation anomaly in the West Pacific (online supplementary figure S1) is prominent in figure 4 (upper panel). In the Atlantic sector there is an apparent intensification of rainfall within the Atlantic Ocean Inter-Tropical Convergence Zone (ITCZ), and a hint of increased rainfall over the Amazon basin. Wetter-than-normal conditions across Amazonia were reported by Marengo and Espinoza (2016), including unprecedented rainfall in western Amazonia (Espinoza *et al* 2014). The anomalous 200 hPa divergence (second panel) in the reanalysis shows features in some of the same places as the rainfall. The relative intensities of the Pacific features and those over the Atlantic sector are more closely balanced in the divergence field than they are in the rainfall. In particular, over tropical South America the divergence anomalies appear more pronounced than the precipitation anomalies.

Theoretical understanding of Rossby wave forcing has provided the concept of a Rossby wave source diagnostic (RWS; Sardeshmukh and Hoskins 1998, James 1994). The RWS is the rate of change in absolute vorticity of an air parcel following the non-divergent part of the flow, and is composed of a vertical stretching term (the product of the divergence and absolute vorticity) and the advection of absolute vorticity by the divergent part of the flow. As a result, divergence anomalies in the deep tropics are ineffective as direct Rossby wave sources because the absolute vorticity, and vorticity gradients, are small there (second panel in figure 4). In addition to the anomalous divergence over tropical South America there are anomalies to the north over the Atlantic which are in regions of higher absolute vorticity and vorticity gradients and which therefore could act as sources. Northward flow from the Amazon region intersects the meridional absolute vorticity gradient

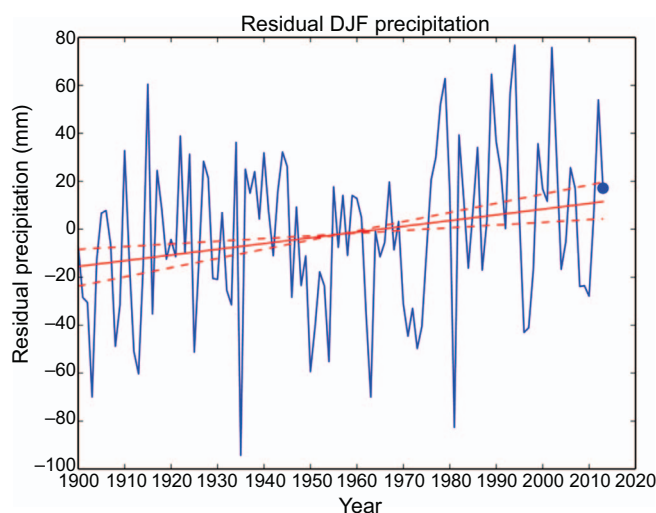


Figure 5. Residual England and Wales (EWP) mean winter precipitation after removal of circulation effects for 1900–01 to 2013–14. The red line is a best-fit linear trend to the residual data and the dashed lines represent the 90% range of statistical uncertainty in the trend.

over the northern coast of South America (third panel in figure 4), which would act to create a negative vorticity (anticyclonic) tendency. In general, however, the anomalous RWS (bottom panel) over the tropical North Atlantic is dominated by the stretching term (not shown), following a pattern that clearly shows its relationship with divergence anomalies. We also reproduce the streamfunction anomalies from the observational analysis in figure 3 to show the relationship with the dynamical response. The large cyclonic streamfunction anomaly in the tropical Atlantic, which appears to be the start of the North Atlantic wavetrain shown in figure 3, is just downstream from the positive RWS anomaly (i.e. source of anomalous cyclonic vorticity) over the Caribbean. This suggests that air parcels moving in the westerly flow in this region acquire an anomalous cyclonic tendency. Anomalous convergence over the Caribbean, therefore, was likely to be the source of the Rossby wave that contributed to the large circulation anomalies in the North Atlantic. In turn, this feature appears to be linked to the divergent flow from South America, suggesting enhanced Amazonian convection could be considered to be the physical driver of the Rossby wave. Note also that there are further positive RWS anomalies to the south of the equatorial divergence associated with the start of a symmetrical wavetrain in the Southern Hemisphere, which is also consistent with the central role of the equatorial anomalies. This feature was associated with a particularly marked drought in southern Brazil (Coelho *et al* 2016).

4. The contribution of climate change

As discussed in the introduction, climate change could have contributed to the extreme UK winter precipitation of 2013–14 either through a dynamical effect on

the atmospheric circulation or through a thermodynamic effect on the moisture capacity of the atmosphere. The former is very complex to assess, as shown by the wide range of model projections of future European rainfall (Collins *et al* 2013, Fereday *et al* 2017) and sensitivity to model formulation (Scaife *et al* 2012). We can address the latter mechanism, however, by comparing observed precipitation with the precipitation expected from the observed circulation. To quantify past circulation, we categorise daily MSLP fields in the region surrounding the UK (0° – 10° W, 50° – 60° N) into 30 different weather types using *k*-means cluster analysis (Fereday *et al* 2008, Neal *et al* 2016). This gives a classification in which each day in the period 1900 to 2014 is assigned to a single weather type. Mean DJF EWP is then reconstructed by randomly assigning each day in the season with a EWP value from a day with the same circulation type in the same month but from a different year between 1931–32 and 2013–14 (for which daily EWP data is available). Random assignment allows 100 alternative seasonal reconstructions to be produced. An estimate of the residual contribution to precipitation variability that is not related to circulation is then computed by subtracting the median reconstructed precipitation value from the EWP series. The reconstruction technique is highly effective, producing winter mean precipitation values which have a correlation of 0.88 with observed precipitation. The standard deviation of the residual precipitation (34 mm) is considerably smaller than that of the reconstructed precipitation (64 mm).

The residual England and Wales winter precipitation is shown in figure 5, and shows that winter 2013–14 was not exceptional by this measure. This emphasises that the record precipitation was related mainly to unusual patterns of atmospheric circulation.

The residual precipitation has considerable year-to-year variability, related to factors such as temperatures in the source regions of air masses transported to the UK. The best-fit linear trend is 0.24 mm yr^{-1} over the 1900–2013 period, which is significant at the 5% confidence level. By contrast, the observed winter EWP for the same period has a trend of 0.18 mm yr^{-1} , which is not significant. This implies that there has been a systematic increase of about 27 mm (about 8% of the long-term average) since 1900 in winter precipitation that is not a result of changing circulation patterns. While it is not possible to uniquely attribute this long-term change without further analysis, it is approximately consistent with the $7\% \text{ K}^{-1}$ increase in potential atmospheric moisture content and the observed increase in annual UK temperatures (Parker and Horton 2005). Increases in global and regional temperatures have been attributed to anthropogenic forcing (IPCC 2013), so it is likely that the residual precipitation trend represents a permanent increase in EWP in response to climate change. Despite this, the size of the implied effect is only a small fraction of the EWP anomaly observed in winter 2013–14 (circa 200 mm), showing that atmospheric circulation variability remains key to explaining the extreme rainfall. Of course, we have not evaluated the degree to which climate change has altered the likelihood of the types of circulation patterns seen in winter 2013–14, which could substantially increase its contribution to the observed extreme rainfall.

5. Conclusions

Record rainfall in the UK in winter 2013–14 resulted from an extreme cyclonic anomaly over the North East Atlantic Ocean that was part of a disrupted pattern of atmospheric circulation across the Northern Hemisphere. Experiments with two atmosphere–ocean models, in which parts of the atmospheric domain are constrained to be similar to the atmospheric state observed in winter 2013–14, are used to diagnose possible global influences on the Northern Hemisphere circulation. These show that the anomalously strong stratospheric polar vortex, likely related to the westerly phase of the QBO, would have favoured a positive phase of the NAO. Relaxation across the full width of the tropical band allows much of the pattern of observed northern hemisphere MSLP anomalies to be simulated, including the deep cyclonic anomaly over the north eastern North Atlantic. This shows that the unusual circulation (and thus the extreme UK conditions) were indeed strongly influenced by the state of the tropical atmosphere. Further experiments reveal that conditions in the Atlantic sector of the tropics contributed to the overall response. Although the North Atlantic responses in both models are weaker than observed, the simulations are capable of

reproducing both of the circulation features that are necessary to explain the severity of the 2013–14 UK weather conditions, namely the north to south gradient of anomalous North Atlantic MSLP and southward displacement of the pattern in the eastern North Atlantic.

Analysis of the upper troposphere in the experiments reveals that the surface responses are related to a Rossby wave train propagating from the tropical Atlantic into the North Atlantic. The dynamical sources of the tropical Rossby wave forcing are shown to be anomalous upper-level convergence over the Caribbean related to divergence over the Amazon basin, corresponding to an intensification of the local Hadley circulation. Further discussion of the dynamical analysis is provided in the online supplementary information. We specifically note, however, that while our analysis focuses on seasonal mean anomalies, examination of events within the season (Davies 2015) may provide additional insight into the causes of the extreme rainfall.

We lastly test the role of climate change on producing the record rainfall seen in winter 2013–14 by inferring long term trends in the part of the rainfall variability unrelated to atmospheric circulation. The residual, non-circulation component of EWP data shows a significant upward trend that corresponds to an increase of 27 mm (circa 8%) of winter mean precipitation since 1900, approximately consistent with the observed warming of the climate. While this implies that climate change contributed to the record rainfall in 2013–14, the size of this contribution is relatively small compared to the overall seasonal precipitation anomaly. The extreme winter rainfall is, therefore, mostly attributable to atmospheric circulation variability. This conclusion could be modified substantially, however, if climate change has already altered the atmospheric circulation. Finally, we note that even a small shift in mean climate could make return periods of events like winter 2013–14 substantially less than in the pre-industrial climate (Schaller *et al* 2016).

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