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LETTER

Tropical origin for the impacts of the Atlantic Multidecadal Variability on the Euro-Atlantic climate

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Abstract

Atlantic Multidecadal Variability (AMV) is known for influencing the mid-latitude climate variability, especially over the European region. This letter assesses the impact of the wintertime AMV in a group of 200-year atmospheric-only numerical experiments, in which the atmosphere is forced with positive and negative AMV-like sea surface temperatures (SSTs) and sea ice concentration patterns. Anomalies are applied separately to the whole North Atlantic ocean, to the extratropics (north of 30° N) and to the tropics (between 0° and 30° N). Results show that AMV anomalies considerably affect the North Atlantic Oscillation (NAO), the jet stream variability and the frequency of atmospheric blocking over the Euro-Atlantic sector, resulting in a negative (positive) NAO during positive (negative) AMV. It is found that the bulk of the signal is originated in the tropics and it is associated with a Gill-like response—an anomalous upper tropospheric streamfunction dipole over the tropical Atlantic driven by the SST anomalies—and with the subsequent structural change of the upper-tropospheric jet, which affects the propagation of Rossby waves in the North Atlantic. Conversely, the NAO response is almost negligible when the AMV anomalies are applied only to the extratropics, suggesting that the relevance of SST anomalies along the North Atlantic frontal zone may be overestimated.

1. Introduction

The Atlantic Multidecadal Variability (AMV), also known as Atlantic Multidecadal Oscillation (AMO), is a long-term alternation of warm and cold decades of sea surface temperatures (SSTs) in the North Atlantic (Schlesinger and Ramankutty 1994, Kerr 2000). AMV can have large impacts on rainfall, temperature and pressure in many regions of the Northern Hemisphere (Knight *et al* 2006), influencing the Atlantic Hurricanes intensity and frequency (Trenberth and Shea 2006), the Sahelian and the Indian/Asian Monsoons (Lu *et al* 2006, Zhang and Delworth 2006, Kucharski *et al* 2009) and even the mid-latitude climate (Häkkinen *et al* 2011).

The linkage between the AMV and the Euro-Atlantic climate is not totally understood: due to the large internal variability of the mid-latitude atmosphere and to the few cycles of the AMV in the observational period, it is hard to identify a robust connection between the two, especially in terms of causality. A common way to describe the mid-latitude atmospheric variability involves the North Atlantic Oscillation (NAO). The NAO is a mode of variability characterized by an oscillation of the pressure gradient between high and mid-latitudes over the North Atlantic (Hurrell *et al* 2003): during the positive phase the eddy-driven jet is shifted poleward and is well separated from the subtropical jet, whereas during the negative phase the eddy-driven jet is shifted equatorward and almost merged with the subtropical jet. The NAO pattern is often identified via Empirical Orthogonal Functions (EOFs) analysis (Ambaum *et al* 2001), but can also be characterized by the frequency of occurrence of Euro-Atlantic weather regimes (Cassou *et al* 2004, Ferranti *et al* 2014).

In the last few years the NAO phases and the jet variability have been described in terms of Rossby

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Wave Breaking (RWB) events, defined as the reversal of the potential temperature gradient measured at the tropopause level (McIntyre and Palmer 1983). RWB can be classified into cyclonic/anticyclonic events according to the environmental meridional shear of the zonal wind, usually known as barotropic shear (Thorncroft *et al* 1993, Peters and Waugh 1996, Tyrilis and Hoskins 2008). Cyclonic wave breaking is dominant on the poleward side of the jet-stream (i.e. over Greenland), whereas anticyclonic wave breaking is usually detected on the equatorward side (i.e. over Eastern Atlantic and Western Europe). Successive RWB events modulate the sign of the NAO and consequently the position of the jet stream (Benedict *et al* 2004, Franzke *et al* 2004, Strong and Magnusdottir 2008). Long-lasting RWB events are often characterized by atmospheric blocking (Pelly and Hoskins 2003), a persistent and quasi-stationary high-pressure system frequently occurring during wintertime at the exit region of the jet stream (Rex 1950, Davini *et al* 2012). Over the Euro-Atlantic region atmospheric blocking is intimately linked with the eddy-driven jet stream variability (Woollings *et al* 2010a) and with North Atlantic Oscillation phases (Woollings *et al* 2008). Indeed, high-latitude blocking over Greenland has been shown to be negatively correlated with the NAO, suggesting that the NAO itself could be interpreted as series of successive blocked and non-blocked conditions (Woollings *et al* 2010b). Since they are dynamically-based, measures of atmospheric blocking—and especially recent 2-dimensional indices which evaluate also RWB properties (Davini *et al* 2012, Masato *et al* 2012)—represent ideal diagnostics to investigate the mid-latitude climate variability in Global Climate Models (GCMs) (e.g. Davini and Cagnazzo 2014).

A few works have identified an opposite relation between the North Atlantic SSTs and the NAO phases (Peings and Magnusdottir 2014, Omrani *et al* 2014, 2015, Gastineau and Frankignoul 2015), with positive AMV leading to negative NAO. Recently Peings and Magnusdottir (2014) published in this journal the results of a comprehensive analysis of the relationship between the North Atlantic multidecadal fluctuations of SSTs and the insurgence of specific Euro-Atlantic weather regimes. They concluded that, in both reanalysis and GCM simulations, the positive phase of the AMV leads to an increased frequency of the NAO- and blocking regimes. Furthermore, positive AMV results in higher frequencies of cold spells over Europe and Eastern United States, in accordance with what has been recorded in the last few years (e.g. Coumou and Rahmstorf 2012).

In the present letter we start from the work of Peings and Magnusdottir (2014) to further investigate the relationships existing between the AMV and the mid-latitude climate. We make use of a group of long atmosphere-only experiments with a state-of-the-art high-resolution GCM and of a diagnostic based on the frequency of atmospheric blocking (Davini *et al* 2012). More specifically, we investigate the different contributions of the AMV to the Euro-Atlantic variability, distinguishing the signal generated by extratropical SSTs from that associated with tropical SSTs. This aims at providing a comprehensive assessment of the impacts of the AMV on the winter mid-latitude atmosphere and on the relevant dynamical mechanisms in action.

2. Data and method

We use the atmospheric component of version 3.1 of the Earth-System Model EC-Earth (Hazeleger *et al* 2010), composed of the ECMWF Integrated Forecast System (IFS, cycle 36r4, ECWMF 2009) atmospheric component with the the H-TESSEL land surface scheme (Balsamo *et al* 2009)

With respect to the previous version (v3.0.1), EC-Earth 3.1 shows a reduced radiative imbalance and an improved hydrological cycle. The atmospheric configuration of IFS adopted (T255L91) is equivalent to a horizontal resolution of about 80 km. There are 91 vertical levels with the last full model level at 1 hPa.

We perform a group of atmosphere-only simulations with prescribed oceanic boundary conditions applying different anomalies representing the positive and negative phases of the AMV. Sea Surface Temperatures (SSTs) and Sea Ice Concentrations (SICs) data from the HadISST dataset (Rayner *et al* 2003) have been used.

We first define the NASST index as the area-averaged yearly anomalies of the North Atlantic SSTs between 75° W–5° W and 0° −70° N from 1870 up to 2012. In analogy with other works (Trenberth and Shea 2006, Peings and Magnusdottir 2014), the Atlantic Multidecadal Variability (AMV) index is then constructed by subtracting from the NASST index the 10-year running mean of the global SSTs (area-averaged between 60° S and 60° N). This is done in order to avoid the effect of the recent global warming.

Afterwards we identify years of positive/negative AMV extracting the upper/lower quartiles of the distribution of the AMV index (defined over the 1870–2012 period). SST and SIC anomalies for the chosen years are then averaged on monthly basis in order to create an ideal seasonal cycle for both positive and negative AMV phases. Even though sea-ice is set to two different climatological values before 1900 and between 1940–1952 (Rayner *et al* 2003), and thus its variability can be underestimated, we decided to use the full HadISST dataset in order to sample the largest possible variability for SST. We did not further detrend the anomalies in order to keep the anomalies more uniform and realistic, although this method could potentially underestimate the effect of global warming at high latitudes (Ting *et al* 2011).

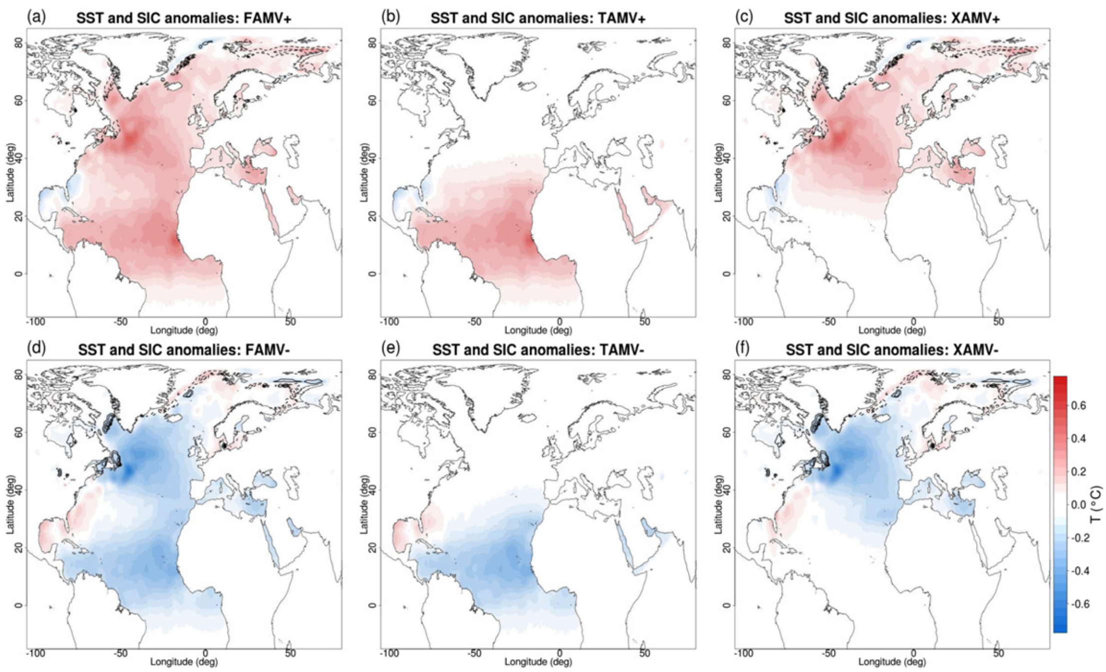

Figure 1. Sea surface temperature (colors) and sea ice concentration (contours) anomalies applied in the different experiments. For SIC, contours are drawn every 2%, solid is positive and dashed is negative.

The way in which we superimpose such ideal anomalies on the climatological seasonal cycle of the HadISST dataset for the 1990–2010 period defines 6 different experiments: FAMV+ and FAMV− (Full AMV) are two experiments where the (positive and negative, respectively) anomalies are applied to the whole North Atlantic and the Mediterranean Sea, from the equator up to the Arctic Ocean. TAMV+ and TAMV− (Tropical AMV) are two experiments in which the anomalies are applied only between 0° N and 30° N. Finally, XAMV+ and XAMV− (Extra-tropical AMV) are two experiments where the anomalies are applied north of 30° N. On the edges of the anomalies (e.g. at the equator for FAMV), an exponential smoothing is applied to avoid the presence of an artificial step in SST. The winter averages (December to March) of the 6 different SST and SIC anomalies are reported in figure 1. A supplementary simulation with no anomalies was run as a reference experiment: hereafter it will be named CONTROL.

Well mixed greenhouse gases, stratospheric ozone and volcanic aerosol concentrations have been fixed at the values of year 2000 for all simulations, according to the historical scenario of the Coupled Model Intercomparison Project—Phase 5 (CMIP5) protocol (Taylor *et al* 2012).

The 7 simulations are run for a total of 201 years each. Given the small time needed for atmosphere and land surface to adjust to the oceanic boundary conditions, we discard the first 6 years of simulation: all the following figures are referring to simulation years 7–201 (195 years).

In order to evaluate biases of the EC-Earth simulations we use the ECMWF ERA-INTERIM Reanalysis (Simmons *et al* 2007) from year 1979 up to 2015. Since we want to study the AMV impact on the European climate, we focus on the season in which the mid-latitude climate variability is more pronounced, i.e. wintertime (defined as December to March, DJFM).

As mentioned in the introduction, a powerful diagnostic to evaluate the properties of the mid-latitude climate variability is provided by atmospheric blocking. To objectively recognize blocking events a 2 d index based onto the reversal of the gradient of geopotential height measured at 500hPa (Z500) has been adopted.

Firstly, both EC-Earth and ERA-INTERIM 500hPa geopotential height fields are interpolated on a common 2.5° x 2.5° regular grid with a second order conservative remapping. Then, as in Tibaldi and Molteni (1990), two meridional gradients of geopotential height are defined:

$$GHGS\left(\lambda_0, \phi_0\right) = \frac{Z500\left(\lambda_0, \phi_0\right) - Z500\left(\lambda_0, \phi_S\right)}{\phi_0 - \phi_S}, \tag{1}$$

$$GHGN\left(\lambda_0, \phi_0\right) = \frac{Z500\left(\lambda_0, \phi_N\right) - Z500\left(\lambda_0, \phi_0\right)}{\phi_N - \phi_0} \tag{2}$$

but here ϕ_0 ranges from 30° N to 75° N and λ_0 ranges from 0° to 360°. $\phi_S = \phi_0 - 15°$, $\phi_N = \phi_0 + 15°$. Instantaneous blocking is thus identified when:

$$GHGS\left(\lambda_0, \phi_0\right) > 0$$
$$GHGN\left(\lambda_0, \phi_0\right) < -10\text{m}/^{\circ}\text{lat} \tag{3}$$

Further constraints have been applied to instantaneous blocking. Firstly, large scale blocking is defined when an instantaneous blocking is extended for at least 15° of continuous longitude. Secondly, a large scale blocking event is defined for each grid point when a large scale blocking occurs within 5° lon (2 grid points) and 2.5° lat (1 grid point) of it. Finally, a blocking event at a certain grid point is defined when a large scale blocking event lasts for at least 5 days. Those constraints ensure that blocking events have a significant longitudinal extension, are persistent and quasi-stationary. The percentage of days per season in which blocked events occur (i.e. blocked days) defines the blocking climatology. A complete description of the blocking detection scheme may be found in Davini *et al* (2012).

We make use of other diagnostics to identify the properties of the signal at the mid-latitudes. We use the Eady growth rate maximum, defined following Vallis (2006):

$$\sigma_{BI} = 0.3068 f \left| \frac{\partial U(z)}{\partial z} \right| N^{-1} \tag{4}$$

where $U(z)$ is the vertical profile of the eastward wind component, f the Coriolis frequency and N is the Brunt–Väisälä frequency, defined as:

$$N^2 = \frac{g}{\theta}\frac{\partial \theta}{\partial z} \tag{5}$$

where θ is the air potential temperature and g the standard gravity acceleration. The impact of the transient eddies is also evaluated computing the transient eddy activity as the bandpass filtered (2–6 days) standard deviation of the 500-hPa daily geopotential height. This also provides an indirect measure of the position of the North Atlantic storm track.

Finally, a 1000-trial bootstrap method is used to evaluate the level of significance of the mean differences between the simulations. Bootstrap has been chosen considering the non-Gaussianity of the blocking distribution. To test against the null hypothesis of no difference between two experiments, we artificially create a set of 1000 differences—computed subtracting the average of two random subsets of 195 years each, obtained randomly sampling the original yearly averaged data of the two experiments. The original difference between the two experiments is then compared point by point with the distribution of the differences obtained from the randomly-shuffled subsets. The level of significance is fixed at 2% if not stated differently (two-sided test).

3. Results

3.1. Blocking sensitivity to AMV anomalies

The representation of atmospheric blocking is known for being a common issue in GCMs: for instance, a large systematic negative bias over Europe has been reported by both atmospheric (d'Andrea *et al* 1998, Neale *et al* 2013) and coupled climate models (Anstey *et al* 2013, Masato *et al* 2013).

Figure 2(a) shows the DJFM blocking climatology of the EC-Earth 3.1 CONTROL experiment (contours) and its bias versus the ERA-INTERIM Reanalysis (colors), both expressed as percentage of blocked days. A pattern with three relative maxima over North Pacific, Greenland and Northern Europe is evident, with values around 10–15%. The model shows a limited negative bias over North Pacific and Europe. Such bias is almost negligible over Greenland, which is a key region for the correct representation of the NAO and of the jet stream variability (Woollings *et al* 2008, Davini and Cagnazzo 2014). Conversely, blocking frequency is overestimated at low latitudes over both the Pacific and Atlantic oceans and over Western Russia. Overall, the representation of blocking in EC-Earth 3.1 is good. Given the linkage among blocking, NAO, eddy-driven jet and weather regimes discussed in the introduction, we can conclude that the representation of the main elements of the Euro-Atlantic mid-latitude variability can be considered as fairly realistic.

Considered the quality of the blocking climatology in EC-Earth 3.1, we investigate the effects of the different phases of the AMV on the blocking frequencies. Figure 2(b) reports the FAMV+ minus FAMV− blocking anomaly, providing the relative difference between the positive and the negative FAMV phases. An evident dipole on the two sides of the jet stream is detected, showing an increased blocking activity over Greenland and Northern Europe and a reduction at lower latitudes over Central Atlantic. There is about a 26% relative increase of blocking over Greenland (defined as a box over 65°–15° W; 62.5°–72.5° N) and 14% relative increase over Northern Europe (15° W–25° E; 60°–65° N). A 25% relative decrease is observed over the Central Atlantic (60°–20° E; 30°–40° N). Considered the strong relationship between blocking over the North Atlantic and the NAO (figure S4 in the supplementary material, available at stacks.iop.org/erl/10/094010/mmedia), this implies the presence of an opposite relation between AMV and NAO (i.e. positive AMV resulting in a negative NAO, and vice-versa), which is in agreement with works reporting an increase in blocking activity over the North Atlantic during AMV+ (Häkkinen *et al* 2011) and with works showing a connection between AMV+ and NAO− (Peings and Magnusdottir 2014, Gastineau and Frankignoul 2015).

An almost identical pattern with comparable intensity can be observed when anomalies are applied only to the tropical Atlantic (experiments TAMV+

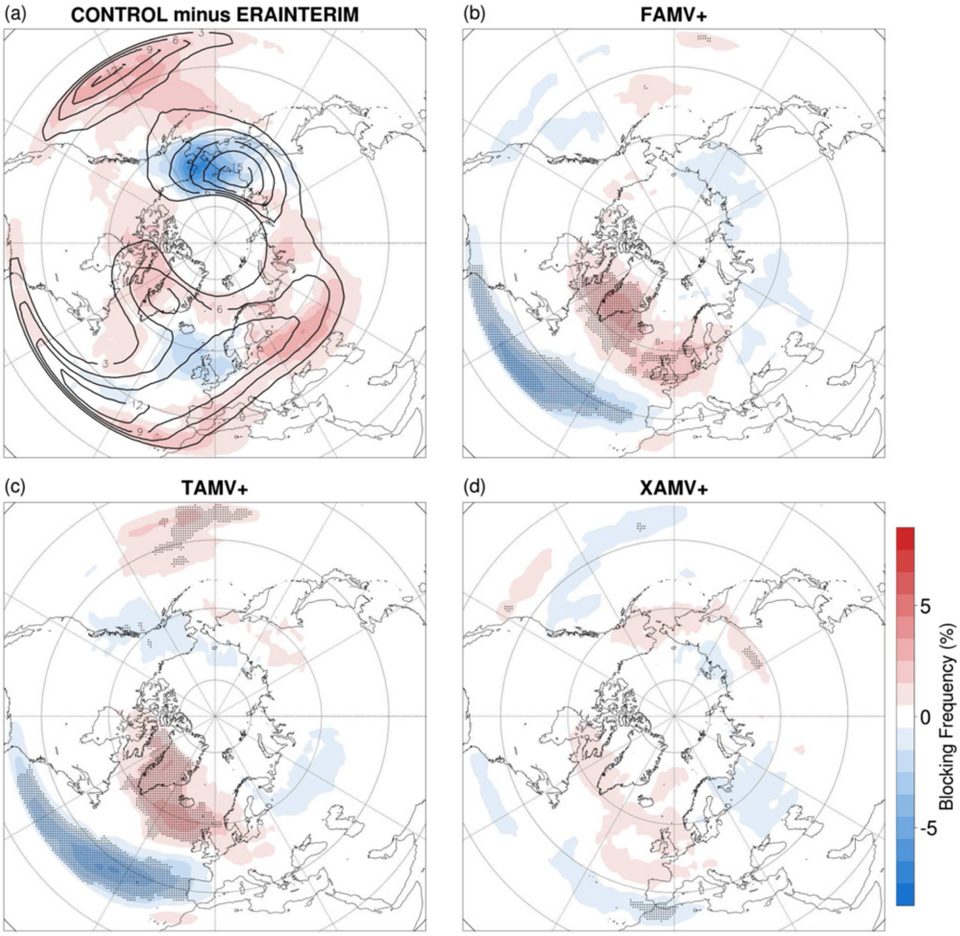

Figure 2. (a) DJFM Blocking frequency bias for the CONTROL experiment with respect to the ERA-INTERIM reanalysis (colors) and CONTROL blocking frequencies (contours). DJFM Blocking frequency anomalies shown as positive minus negative phase for (b) FAMV, (c) TAMV and (d) XAMV experiments. All are expressed as percentage of blocked days per season. In (a) contours are drawn each 3%. Stippled regions show significance at the 2% level.

minus TAMV−, figure 2(c)). Conversely, a barely noticeable signal is observed when SST and SIC anomalies are applied only to the mid-latitudes (figure 2(d), XAMV+ minus XAMV−). We measure a relative increase of about the 6% (3%) over Greenland (Central Europe) and only a 3% decrease over Central Atlantic. None of these anomalies is statistical significant even at the 10% level. This indeed suggests that the observed effects of the Atlantic Multidecadal Variability in EC-Earth 3.1 are driven by the tropics.

The seasonality of the response (DJ vs FM) is the only feature where we can find some differences between FAMV and TAMV experiments (supplementary material figures S1 and S2). FAMV experiments shows a stronger response in early winter (DJ), while TAMV experiments shows stronger response in later winter (FM).

We also investigated the linearity of the response for FAMV and TAMV, looking at the difference between each experiments and the control run (supplementary material figure S3). The response is surprisingly linear, showing the negative/positive NAO-like pattern for positive/negative AMV anomalies. A minor difference can be observed over Scandinavia: indeed here we note a weak increase of blocking in both FAMV+/TAMV+ and FAMV−/TAMV−.

3.2. Large-scale changes

In order to detect the origin of this tropical signal, we investigate the large-scale changes associated with the AMV anomalies. In figure 3 DJFM changes of the 2 m temperature, of sea level pressure, of precipitation and of the 300-hPa streamfunction are shown.

Surface temperature anomalies are related to the SST patterns imposed in the different experiments: the low-level westerly flow advects warmer oceanic air from the North Atlantic downstream up to Eurasia, leading to an overall warming or cooling (Seager *et al* 2002) according to the sign of the Atlantic SST anomalies.

Sea level pressure reflects the changes already seen in atmospheric blocking, with TAMV and FAMV experiments showing a NAO-like dipole over the North Atlantic, and XAMV showing a weak signal over Greenland.

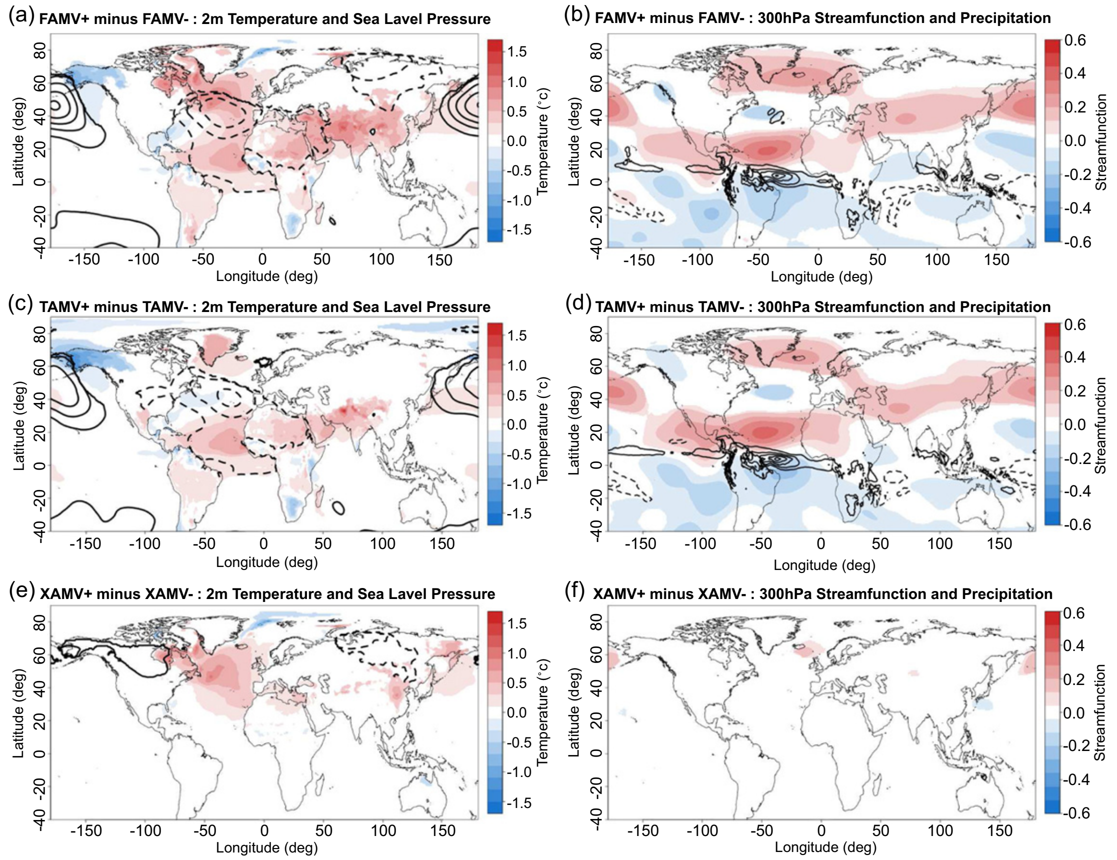

Figure 3. Anomalies of 2 m temperature (color) and sea level pressure (SLP, contours) for (a) FAMV, (c) TAMV and (e) XAMV experiments. Anomalies of 300-hPa streamfunction (colors) and precipitation (contours) for (b) FAMV, (d) TAMV and (f) XAMV experiments. Anomalies are expressed as positive minus negative AMV phase. Solid contours is positive and dashed is negative. For SLP, contours are drawn each 0.5 hPa. For precipitation, contours are drawn each 0.5 mm day^{-1}. Only values where the 2% significant level is reached are drawn.

Interestingly, we observe an increase in precipitation a few degrees north of the equator for FAMV+ minus FAMV− and TAMV+ minus TAMV−. This signal is consistent with the warmer SST anomalies imposed, which lead to increased evaporation and convection. This is confirmed by the outgoing long-wave radiation anomalies in the same region (not shown). Conversely, XAMV does not show any evident precipitation anomaly.

In addition, both FAMV and TAMV experiments show an equatorial dipole in the upper-tropospheric streamfunction, characterized by two anticyclonic anomalies at 300 hPa across the equator at 20° N and 10° S (figures 3(b)–(d)). This implies a change in the zonal winds over the Tropical Atlantic, with stronger westerlies right north of the streamfunction anomaly. It is interesting to note that the the diabatic heat source associated with the precipitation anomaly is placed exactly on the node of the 300-hPa streamfunction dipole, suggesting a possible dynamical relationship between the two.

A final note should be devoted to the weakening of the Aleutian low observed in both FAMV+ and TAMV+. This is likely due to the excitement of Rossby Wave trains associated with the minor poleward displacement of the ITCZ over the Pacific (e.g. Okumura *et al* 2009)—which in turn may be due to changes in the Walker circulation. However, a deeper analysis of the signal over the Pacific goes beyond the scope of this study.

3.3. The mechanism: the Gill response and the shift of the subtropical jet

In their GCM-based work, Sutton and Hodson (2007) applied a positive SST anomaly to the Tropical Atlantic and found a dipolar anticyclonic response at upper levels centered around the equator. This anticyclonic signal was accompanied by a cyclonic anomalies at lower levels. They highlighted that this response shared many features with the linear model presented by Gill (1980), where an off-equatorial heating produces a stationary equatorial Rossby wave to the north-western and south-western side of the heating source, characterized by a single baroclinic mode. Even though Sutton and Hodson (2007) underlined that the strongest response was produced in summer, they remarked that the mechanism was solid throughout all seasons.

Interestingly, the precipitation and streamfunction anomalies discussed in the previous section are in agreement with the results obtained by Sutton and Hodson (2007), suggesting that the streamfunction dipole is produced by the latent heat release driven by the increased precipitation.

This agrees with the idea that in FAMV and TAMV experiments we observe a Gill-like response, with an upper level anticyclone and a weaker lower level cyclone (not shown) when a warm SST anomaly is imposed. The opposite occurs with a negative anomaly. This is further supported by the negative SLP anomalies seen over Central and Northern Africa (figures 3(a)–(c)), possibly a combination of the Kelvin part of the Gill-like solution and of the negative NAO signal over the North Atlantic.

As may be expected, this signal is stronger in summer (not shown). Furthermore, the fact that the tropical streamfunction anomaly is not observed in the XAMV experiments confirms the critical role of the tropical precipitation anomalies.

But how do the tropical streamfunction anomalies propagate at higher latitudes and affect the Euro-Atlantic blocking frequency? As mentioned in the introduction, atmospheric blocking and eddy-driven jet stream variability are strongly associated with Rossby Wave Breaking events. The positive feedback between baroclinic eddies and the mean flow is fundamental for the existence of the latitudinal variability of the North Atlantic eddy-driven jet stream (Limpasuvan and Hartmann 2000, Hartmann 2007). This means that the mean state of the jet stream can affect the main features and the frequencies of cyclonic and anticyclonic wave breaking, which in turn can affect the position of the jet via the eddy-mean flow feedback. Numerical simulations showed that the more north/south the mean jet is placed, larger frequencies of anticyclonic/cyclonic Rossby Wave Breaking are counted, in both simplified model (Barnes and Hartmann 2012) or GCM experiments (Barnes and Polvani 2013, Davini and Cagnazzo 2014). A key role is played by the environmental barotropic wind shear in which RWB occurs (Thorncroft *et al* 1993, Peters and Waugh 1996). Rossby waves propagating in a cyclonic barotropic shear (i.e. north of the jet stream) will tend to break cyclonically and poleward, whereas equatorward anticyclonic wave breaking will be favored south of the jet (Hartmann and Zuercher 1998, Tyrilis and Hoskins 2008). Once the breaking of the wave has occurred, this in turns affects the jet stream structure over the North Atlantic and may lead to the formation of cut-off lows or high-pressure anomalies (i.e. blocking).

A direct consequence of the 300-hPa streamfunction anomaly shown in figure 3 is the increase of the zonal winds between 20° N and 40° N, on the southern edge of the subtropical jet stream over the Atlantic. This is shown in figure 4(a): over Eastern North America (between 90° W and 50° W) there is a single jet stream around 40° N, and it is shifting equatorward in FAMV+ by about 2° with respect to FAMV−. This implies a change in the environmental barotropic wind shear in a region where Rossby waves propagate and start their breaking.

Figure 4(b) shows the climatological cross-section of the zonally averaged meridional wind shear, while the FAMV+ minus FAMV− field is reported in colors. The latter is characterized by a negative anomaly right where the maximum of the jet stream is found. In terms of Rossby wave dynamics, a negative wind shear anomaly (which is climatologically observed poleward of the eddy-driven jet stream) will favor the occurrence of cyclonic wave breaking events. Cyclonic wave breaking is tightly coupled with the occurrence of Greenland blocking and with the negative phase of the North Atlantic Oscillation (Pelly and Hoskins 2003, Woollings *et al* 2008), just observed in figure 2(b).

The shift of the jet is reflected by the anomalies of the Eady growth rate maximum, which shows an equatorward shift of the upper tropospheric baroclinicity (figure 4(c)). As a direct consequence of the eddy-mean flow feedback, an equatorward migration of the transient eddies activity is observed as well (figure 4(d)). The same patterns are obtained for TAMV+ minus TAMV− (figure S5 in the supplementary material), providing further evidence on the tropical origin of the observed NAO-like signal.

This mechanism shares many features with the one identified by Terray and Cassou (2002), in which positive tropical SST anomalies lead to increased baroclinicity and a shift in the storm track driven by planetary wave changes.

Conversely, those structural changes of the jet stream are not observed in the XAMV+ minus XAMV − case. Here we observe a minor reduction of the Eady Growth Rate maximum in the lower troposphere in proximity of the eddy-driven jet stream, associated with a decrease of the transient eddies activity (Figure S6 in the supplementary material). As pointed out by Peings and Magnusdottir (2014), such reduced baroclinicity can be connected with the weaker gradient along the SST frontal zone across the Gulf Stream, caused by the positive extratropical SST anomalies imposed in the XAMV+ and FAMV+ simulations. This can influence the properties of the North Atlantic storm track (Brayshaw *et al* 2008) and the RWB properties (Rivère 2009), favoring the occurrence of cyclonic events. However, this mechanism appears negligible if compared to the tropical-induced one observed in both TAMV and FAMV experiments.

4. Discussion and conclusions

In this work we established the presence of a strong connection between winter climate variability over the Euro-Atlantic region and the phases of the AMV in a series of 200-year GCMs runs with EC-Earth 3.1. We found that the large part of the signal over Europe due to imposed AMV-like anomalies comes from the

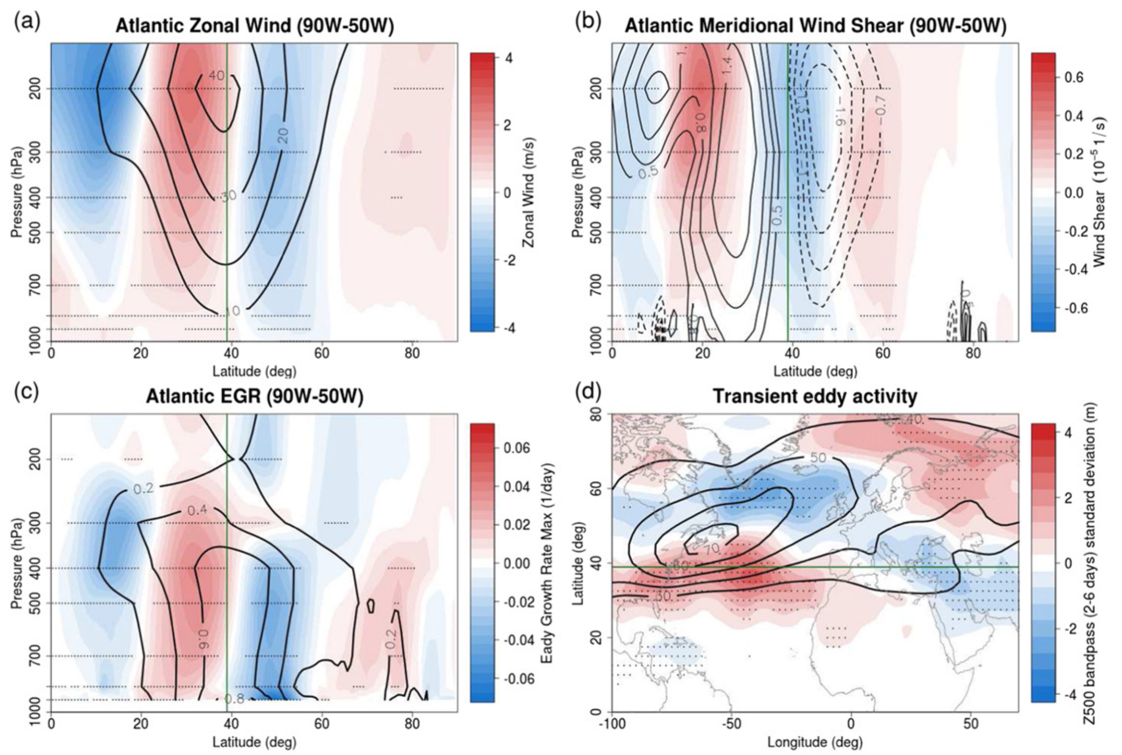

Figure 4. (a) CONTROL cross-section of zonally averaged zonal wind (contours) and FAMV+ minus FAMV− (colors) over the Eastern North America (90W-50W). (b) As (a) but for zonally averaged barotropic wind shear. (c) As (a) but for Eady growth rate maximum. (d) CONTROL transient eddy activity at 500 hPa (contours) and FAMV+ minus FAMV− (colors). Green lines show the position of the maximum of 300-hPa zonal wind averaged between 90W–50W, indicative of the upper tropospheric jet stream. Stippled regions show significance at the 2% level.

tropics, in analogy to what suggested by previous works that highlighted the influence of Tropical Atlantic SSTs on the Euro-Atlantic region (Okumura *et al* 2001, Terray and Cassou 2002, Peng *et al* 2005, Sutton and Hodson 2007). We also discussed two different mechanisms in action, one based in the tropics and one of extratropical origin.

They can be summarized as follows:

– The ‘from above’ mechanism: during the positive AMV phase, higher tropical SSTs enhance convection and precipitation a few degrees north of the equator, leading to a Gill-like response in the Tropical North Atlantic. This affects the subtropical jet stream structure near the Eastern North American coast, adding a negative barotropic wind shear anomaly and shifting the jet maximum to the south. The barotropic wind shear influences the propagation of Rossby waves, increasing the occurrence of cyclonic wave breaking. This ultimately leads to a negative NAO-like signal and a southward displaced eddy-driven jet stream over the North Atlantic, exactly as observed in FAMV+ and TAMV+ experiments. The opposite mechanism, driven by negative SST anomalies, is operating during the negative AMV phase.

– The ‘from below’ mechanism: during positive AMV phases, positive extratropical SST anomalies weaken the meridional gradient of the surface temperatures across the SST frontal zone in proximity of the Gulf Stream. This in turns affects the baroclinicity of the lower troposphere at the exit region of the eddy-driven jet stream, reducing the jet speed and shifting equatorward the jet stream. Conversely, negative SST anomalies increase the SST gradient and increase the low level baroclinicity. This mechanism appears to be extremely weak in our XAMV experiments (i.e. statistically not significant).

Peings and Magnusdottir (2014) found results similar to our FAMV experiments, for both reanalysis and GCM simulations: they concluded that the origin of the NAO-like signal was associated with changes of the North Atlantic SST gradient (i.e. the ‘from below’ mechanism). However, in a more recent experiment Peings *et al* (2015) performed another series of sensitivity integrations in which they split their forcing in a tropical and an extratropical part. In this configuration, consistently with our results, they found a larger contribution by the tropical component, although the extratropical component was a key factor to reinforce the atmospheric NAO signal. Indeed, in our simulations, the barotropic shear changes caused by the tropical precipitation anomalies are clearly leading the observed blocking pattern over the Euro-Atlantic region. However, the partial non-linearity of the signal and the different response in early and late winter (shown in the supplementary material, available at stacks.iop.org/ERL/10/094010) both suggest the presence of a non-linear interaction between the tropical and extratropical SST anomalies. In any case, some of the differences between the two studies can also be associated with the model characteristics and setup.

As can be seen by comparing figure 5(d) of Peings and Magnusdottir (2014) with our figure 4(c), EC-Earth FAMV and TAMV simulations show slightly stronger Eady growth rate anomalies, suggesting that EC-Earth might be more sensitive to SST anomalies. A first possible explanation of this discrepancy may be rooted in the larger SST anomalies that we applied at the tropical level (a methodological discussion on the reason causing these differences can be found in the supplementary material).

A second and more solid possibility can be related to the different atmospheric mean state over the North Atlantic in the model simulations. A jet stream a few degrees too poleward or too equatorward can have a dynamically different dominant mode of variability (Barnes and Polvani 2013, Davini and Cagnazzo 2014) that can translate in a different sensitivity to a diabatic heat source in the tropics. The different mean state can also explain the modest signal observed in the stratosphere (supplementary material figure S7). Such weak response suggests that the stratosphere in EC-Earth—that is a high-top model—does not play a relevant role for the propagation of the AMV signal over Europe, in contrast of what have been reported by Omrani *et al* (2014).

It is clear that a model response to a boundary forcing can be model dependent. On the other hand, the comparison between TAMV and XAMV simulations provides an unquestionable evidence in favor of the ‘from above’ mechanism. This interpretation is reinforced by the fact that the EC-Earth model—here in atmospheric-only configuration—shows minor biases in the representation of the mid-latitude climate. This also suggests that the influence of the SST anomalies along the North Atlantic frontal zone on the European climate may be overestimated.

Our findings shed further light on the importance of the representation of tropical convection in GCMs, which has been recently pointed as one of the largest open issues of state-of-the-art GCMs (Stevens and Bony 2013). Furthermore, the presence of a clear relationship between tropical convection/precipitation and mid-latitude climate variability adds further evidence to the importance of a correct simulation of equatorial dynamics, which can have wide implications for both seasonal and decadal prediction over Europe (e.g. Cassou 2008).

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References

Ambaum M H, Hoskins B J and Stephenson D B 2001 Arctic oscillation or North Atlantic oscillation? *J. Clim.* **14** 3495–507

Anstey J A *et al* 2013 Multi-model analysis of Northern Hemisphere winter blocking: model biases and the role of resolution *Journal of Geophysical Research: Atmospheres* **118** 3956–71

Balsamo G, Beljaars A, Scipal K, Viterbo P, van den Hurk B, Hirschi M and Betts A K 2009 A revised hydrology for the ECMWF model: verification from field site to terrestrial water storage and impact in the Integrated Forecast System *J. Hydrometeorology* **10** 623–43

Barnes E A and Hartmann D L 2012 Detection of Rossby wave breaking and its response to shifts of the midlatitude jet with climate change *J. Geophys. Res.: Atmos.* **117** D09117

Barnes E A and Polvani L 2013 Response of the midlatitude jets, and of their variability, to increased greenhouse gases in the CMIP5 models *J. Clim.* **26** 7117–35

Benedict J, Lee S and Feldstein S 2004 Synoptic view of the North Atlantic Oscillation *J. Atmos. Sci.* **61** 121–44

Brayshaw D J, Hoskins B and Blackburn M 2008 The storm-track response to idealized sst perturbations in an aquaplanet gcm *J. Atmos. Sci.* **65** 2842–60

Cassou C 2008 Intraseasonal interaction between the madden-julian oscillation and the north atlantic oscillation *Nature* **455** 523–7

Cassou C, Terray L, Hurrell J W and Deser C 2004 North Atlantic winter climate regimes: spatial asymmetry, stationarity with time, and oceanic forcing *J. Clim.* **17** 1055–68

Coumou D and Rahmstorf S 2012 A decade of weather extremes *Nat. Clim. Change* **2** 491–6

d’Andrea F *et al* 1998 Northern hemisphere atmospheric blocking as simulated by 15 atmospheric general circulation models in the period 1979-1988 *Clim. Dyn.* **14** 385–407

Davini P and Cagnazzo C 2014 On the misinterpretation of the North Atlantic Oscillation in CMIP5 models *Clim. Dyn.* **43** 1497–511

Davini P, Cagnazzo C, Gualdi S and Navarra A 2012 Bidimensional diagnostics, variability and trends of Northern Hemisphere blocking *J. Clim.* **25** 6496–509

ECWMF 2009 IFS cycle36r1 (http://www.ecmwf.int/research/ifsdocs/CY36r1/) European Center for Medium Range Forecast

Ferranti L, Corti S and Janousek M 2014 Flow-dependent verification of the ecmwf ensemble over the euro-atlantic sector *Q. J. R. Meteorol. Soc.* **141** 916–24

Franzke C, Lee S and Feldstein S 2004 Is the North Atlantic Oscillation a breaking wave? *J. Atmos. Sci.* **61** 145–60

Gastineau G and Frankignoul C 2015 Influence of the north atlantic sst variability on the atmospheric circulation during the twentieth century *J. Clim.* **28** 1396–416

Gill A 1980 Some simple solutions for heat-induced tropical circulation *Q. J. R. Meteorol. Soc.* **106** 447–62

Häkkinen S, Rhines P B and Worthen D L 2011 Atmospheric blocking and Atlantic multidecadal ocean variability *Science* **334** 655–9

Hartmann D L 2007 The Atmospheric General Circulation and Its Variability (125th Anniversary Issue of the Meteorological Society of Japan) *J. Meteorol. Soc. Japan Ser. II* **85** 123–43

Hartmann D L and Zuercher P 1998 Response of baroclinic life cycles to barotropic shear *J. Atmos. Sci.* **55** 297–313
Hazeleger W *et al* 2010 EC-Earth: a seamless Earth-system prediction approach in action *Bull. Am. Meteorol. Soc.* **91** 1357–63
Hurrell J W, Kushnir Y, Ottersen G and Visbeck M 2003 *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* vol 134 ((American Geophysical Union)
Kerr R A 2000 A North Atlantic climate pacemaker for the centuries *Science* **288** 1984–5
Knight J R, Folland C K and Scaife A A 2006 Climate impacts of the Atlantic multidecadal oscillation *Geophys. Res. Lett.* **33** L17706
Kucharski F, Bracco A, Yoo J, Tompkins A, Feudale L, Ruti P and Dell'Aquila A 2009 A Gill-Matsuno-type mechanism explains the tropical Atlantic influence on African and Indian monsoon rainfall *Q. J. R. Meteorol. Soc.* **135** 569–79
Limpasuvan V and Hartmann D L 2000 Wave-maintained annular modes of climate variability *J. Clim.* **13** 4414–29
Lu R, Dong B and Ding H 2006 Impact of the Atlantic Multidecadal Oscillation on the Asian summer monsoon *Geophys. Res. Lett.* **33** L24701
Masato G, Hoskins B and Woollings T J 2012 Wave-breaking characteristics of midlatitude blocking *Q. J. R. Meteorol. Soc.* **138** 1285–96
Masato G, Hoskins B J and Woollings T 2013 Winter and summer Northern Hemisphere blocking in CMIP5 models *J. Clim.* **26** 7044–59
McIntyre M E and Palmer T 1983 Breaking planetary waves in the stratosphere *Nature* **305** 593–600
Neale R B, Richter J, Park S, Lauritzen P H, Vavrus S J, Rasch P J and Zhang M 2013 The mean climate of the community atmosphere model (cam4) in forced sst and fully coupled experiments *J. Clim.* **26** 5150–68
Okumura Y, Xie S-P, Numaguti A and Tanimoto Y 2001 Tropical atlantic air-sea interaction and its influence on the nao *Geophys. Res. Lett.* **28** 1507–10
Okumura Y M, Deser C, Hu A, Timmermann A and Xie S-P 2009 North pacific climate response to freshwater forcing in the subarctic north atlantic: Oceanic and atmospheric pathways *J. Clim.* **22** 1424–45
Omrani N-E, Bader J, Keenlyside N and Manzini E 2015 Troposphere-stratosphere response to large-scale north atlantic ocean variability in an atmosphere/ocean coupled model *Clim. Dyn.* in press
Omrani N-E, Keenlyside N S, Bader J and Manzini E 2014 Stratosphere key for wintertime atmospheric response to warm atlantic decadal conditions *Clim. Dyn.* **42** 649–63
Peings Y, Simpkins G and Magnusdottir G 2015 Wintertime atmospheric response to Atlantic multidecadal variability: effect of stratospheric representation and ocean-atmosphere coupling, personal communication
Peings Y and Magnusdottir G 2014 Forcing of the wintertime atmospheric circulation by the multidecadal fluctuations of the North Atlantic ocean *Environ. Res. Lett.* **9** 034018
Pelly J and Hoskins B 2003 A new perspective on blocking *J. Atmos. Sci.* **60** 743–55
Peng S, Robinson W A, Li S and Hoerling M P 2005 Tropical atlantic sst forcing of coupled north atlantic seasonal responses *J. Clim.* **18** 480–96
Peters D and Waugh D 1996 Influence of barotropic shear on the poleward advection of upper-tropospheric air *J. Atmos. Sci.* **53** 3013–31
Rayner N, Parker D, Horton E, Folland C, Alexander L, Rowell D, Kent E and Kaplan A 2003 Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century *J. Geophys. Res.* **108** 4407
Rex D 1950 Blocking action in the middle troposphere and its effect upon regional climate: I. An aerological study of blocking action *Tellus* **2** 196–211
Rivière G 2009 Effect of latitudinal variations in low-level baroclinicity on eddy life cycles and upper-tropospheric wave-breaking processes *J. Atmos. Sci.* **66** 1569–92
Schlesinger M E and Ramankutty N 1994 An oscillation in the global climate system of period 65-70 years *Nature* **367** 723–6
Seager R, Battisti D S, Yin J, Gordon N, Naik N, Clement A C and Cane M A 2002 Is the Gulf Stream responsible for Europe's mild winters? *Q. J. R. Meteorol. Soc.* **128** 2563–86
Simmons A, Uppala S M, Dee D and Kobayashi S 2007 ERA-Interim: New ECMWF reanalysis products from 1989 onwards *ECMWF Newsletter* **110** 25–35
Stevens B and Bony S 2013 What are climate models missing *Science* **340** 1053–4
Strong C and Magnusdottir G 2008 Tropospheric Rossby Wave breaking and the NAO/NAM *J. Atmos. Sci.* **65** 2861–76
Sutton R T and Hodson D L 2007 Climate response to basin-scale warming and cooling of the North Atlantic Ocean *J. Clim.* **20** 891–907
Taylor K E, Stouffer R J and Meehl G A 2012 An overview of CMIP5 and the experiment design *Bull. Am. Meteorol. Soc.* **93** 485–98
Terray L and Cassou C 2002 Tropical atlantic sea surface temperature forcing of quasi-decadal climate variability over the north atlantic-european region *J. Clim.* **15** 3170–87
Thorncroft C, Hoskins B and McIntyire M 1993 Two Paradigms of Baroclinic Wave Life-Cycle Behaviour *Q. J. R. Meteorol. Soc.* **119** 17–55
Tibaldi S and Molteni F 1990 On the Operational Predictability of Blocking *Tellus* **42A** 343–65
Ting M, Kushnir Y, Seager R and Li C 2011 Robust features of Atlantic multi-decadal variability and its climate impacts *Geophys. Res. Lett.* **38** L17705
Trenberth K E and Shea D J 2006 Atlantic hurricanes and natural variability in 2005 *Geophys. Res. Lett.* **33** L12704
Tyrilis E and Hoskins B 2008 The morphology of Northern Hemisphere blocking *J. Atmos. Sci.* **65** 1653–65
Vallis G 2006 *Atmospheric and Oceanic Fluid Dynamics: Fundamentals and Large-Scale Circulation* (Cambridge: Cambridge University Press)
Woollings T, Hannachi A and Hoskins B 2010a Variability of the North Atlantic eddy-driven jet stream *Q. J. R. Meteorol. Soc.* **136** 856–68
Woollings T, Hannachi A, Hoskins B and Turner A 2010b A regime view of the North Atlantic Oscillation and its response to anthropogenic forcing *J. Clim.* **23** 1291–307
Woollings T, Hoskins B, Blackburn M and Berrisford P 2008 A new Rossby Wave breaking interpretation of the North Atlantic Oscillation *J. Atmos. Sci.* **65** 609–326
Zhang R and Delworth T L 2006 Impact of Atlantic multidecadal oscillations on India/Sahel rainfall and Atlantic hurricanes *Geophys. Res. Lett.* **33** L17712