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# The influence of the Stratospheric state on North Atlantic Weather Regimes

Andrew J. Charlton-Perez,<sup>a\*</sup>, Laura Ferranti<sup>b</sup>, Robert W. Lee<sup>a,c</sup>

<sup>a</sup>Dept. of Meteorology, Univ. of Reading, Reading, UK

<sup>b</sup>European Centre for Medium Range Weather Forecasts, Reading, UK

<sup>c</sup>NCAS-Climate, Univ. of Reading, Reading, UK

\*Correspondence to: Lyle Building, Univ. of Reading, Whiteknights, Reading, Berks, RG6 6BB, a.j.charlton-perez@reading.ac.uk

Stratosphere-troposphere coupling is often viewed from the perspective of the annular modes and their dynamics. Despite the obvious benefits of this approach, recent work has emphasised the greater tropospheric sensitivity to stratospheric variability in the Atlantic basin than in the Pacific basin. In this study, a new approach to understanding stratosphere-troposphere coupling is proposed, with a focus on the influence of the stratospheric state on North Atlantic weather regimes (during extended winter, November to March). The influence of the strength of the lower stratospheric vortex on four commonly used tropospheric weather regimes is quantified. The negative phase of the North Atlantic Oscillation is most sensitive to the stratospheric state, occurring on 33% of days following weak vortex conditions but on only 5% of days following strong vortex conditions. An opposite and slightly weaker sensitivity is found for the positive phase of the North Atlantic Oscillation and the Atlantic Ridge regime. For the North Atlantic Oscillation regimes, stratospheric conditions change both the probability of remaining in each regime and the probability of transitioning to that regime from others. A logistic regression model is developed to further quantify the sensitivity of tropospheric weather regimes to the lower stratospheric state. The logistic regression model predicts an increase of 40-60% in the probability of transition to the negative phase of the North Atlantic Oscillation for a one standard deviation reduction in the strength of the stratospheric vortex. Similarly it predicts a 10-30% increase in the probability of transition to the positive phase of the North Atlantic Oscillation for a one standard deviation increase in the strength of the stratospheric vortex. The stratosphere-troposphere coupling in the European Centre for Medium Range Weather Forecasts, Integrated Forecasting System model is found to be consistent with the re-analysis data by fitting the same logistic regression model.

*Key Words:* STRATOSPHERE-TROPOSPHERE COUPLING, WEATHER REGIMES

*Received ...*

## 1. Introduction and Motivation

It is now widely accepted that variability in the winter stratosphere is coupled to low-frequency North Atlantic variability on medium-range, sub-seasonal and seasonal timescales during the winter season (e.g. [Tripathi et al. 2015b](#); [Sigmond et al. 2013](#); [Scaife et al. 2016](#)). Stratospheric variability, either Stratospheric Sudden Warming (SSW) events or periods in which the vortex is anomalously strong, typically result in anomalous conditions throughout the lower stratosphere on sub-seasonal timescales (between 30 and 60 days) ([Limpasuvan et al. 2004, 2005](#)). During this time period, the annular mode index in the lower stratosphere and troposphere is, on average, shifted toward negative values during SSW events and positive values during strong vortex events ([Baldwin and Dunkerton 2001](#); [Beerli et al. 2017](#)). Negative values of the annular mode index are associated with a weakened polar night jet in the lower stratosphere and a southward shifted eddy-driven jet in the troposphere. Quantifying and understanding the strength and nature of stratosphere-troposphere coupling has important implications for sub-seasonal prediction for a range of different economic sectors, including renewable energy generation ([Grams et al. 2017](#); [Zubiate et al. 2017](#)).

Although this general behaviour has been known for some time, understanding of the mechanisms by which this connection occurs and the extent to which these mechanisms are captured by numerical models remain elusive. A number of recent reviews ([Kidston](#)

*et al.* 2015; Tripathi *et al.* 2015a) discuss in detail the different mechanisms that have been proposed to explain this effect, with a strong emphasis on the influence of lower stratospheric conditions on upper tropospheric eddy momentum fluxes. Careful consideration of the momentum budget during SSW events (Hitchcock and Simpson 2016) suggests that it is the planetary scale tropospheric eddies which are directly influenced by the lower stratospheric state, with resulting changes to the synoptic-scale eddies simply a consequence of eddy-jet feedbacks in the troposphere, but there remains debate in the literature about the details of the mechanism.

Disentangling the mechanisms which underlie stratosphere-troposphere coupling is hampered by, on one-hand, the limited observational record of coupling events which is available and on the other by the challenge of designing experiments which can cleanly distinguish between different coupling mechanisms. One approach to dealing with the limited observational record is to examine stratosphere-troposphere coupling in numerical models. The approach of Hitchcock and Simpson (2014) which uses artificial stratospheric damping to simulate the same SSW event a large number of times demonstrates that the response to SSW events is strongly concentrated in the North Atlantic sector (see their Fig. 11) and the significant amount of variation in the strength of the tropospheric response that is dependent solely upon antecedent tropospheric conditions (see their Fig. 12). To date, the most comprehensive attempt to design experiments which distinguish between different proposed mechanisms for stratosphere-troposphere coupling are those described by Garfinkel *et al.* (2013). Garfinkel *et al.* also highlight the importance of tropospheric processes in determining stratosphere-troposphere coupling, focussing their attention on arguments associated with tropospheric jet persistence and its links to eddy-jet feedbacks. Garfinkel *et al.* propose a significant non-linearity in the sensitivity of different tropospheric states to stratospheric variability. In their idealised model, tropospheric jets centred at 40° latitude were more than three times more sensitive (in terms of the shift in the tropospheric jet position) to stratospheric perturbations than jets centred at 30° or 50° latitude.

In both of these cases, and in other studies including the original investigations of Baldwin and Dunkerton (1999), event-to-event variability in stratosphere-troposphere coupling is apparent. The key hypothesis of our study, linked to previous work, is that much of the event-to-event variability in stratosphere-troposphere coupling is the result of tropospheric low-frequency variability rather than simply a function of the strength and longevity of the SSW or strong vortex event. To investigate this hypothesis an alternative and complimentary approach to understanding stratosphere-troposphere coupling is adopted, through the use of North-Atlantic weather regimes.

The use of weather regimes to describe low-frequency tropospheric variability is reviewed extensively in Hannachi *et al.* (2017). While there is some debate about the uniqueness and relevance of regime analysis in the North Atlantic, a large number of authors have found this approach to be a useful way to describe low frequency variability and its connection to external drivers. In this context, external refers to other parts of the climate system which vary independently of the extra-tropical North Atlantic atmosphere. Most relevant to this work is the approach taken by Cassou *et al.* (2004) and Cassou (2008) whose studies establish how North Atlantic Sea-Surface Temperatures and the Madden-Julian Oscillation influence the frequency of occurrence of different weather regimes in the North Atlantic and the likelihood of transitions between these regimes. Our study adopts a similar approach, but in this case applied to the influence of the strength of the polar vortex in the lower stratosphere as a proxy for stratospheric variability. A number of studies, most recently Karpechko (2015), have shown that the strength of the polar vortex close to the tropopause has the strongest connection to tropospheric variability (perhaps as this is also the region where stratospheric anomalies are longest lived).

A recent study by Beerli *et al.* (2017) began to examine the response of Atlantic weather regimes to changes in the strength of the stratospheric vortex. In this study, we seek to expand on this approach using a simple model of regime behaviour in the troposphere in order to answer the following two questions:

- Which North Atlantic weather regimes are influenced by the stratospheric polar vortex?
- By how much does the probability of occurrence of North Atlantic weather regimes change in response to typical variations of the stratospheric polar vortex?

In answering these questions, it is necessary to first introduce a simple model of North Atlantic weather regimes (henceforth weather regimes) in order to understand how the stratospheric state modifies their behaviour. Once the simple model is established, a logistic regression model is used to quantify the size of the influence of the stratospheric state on the frequency of occurrence of weather regimes. In the first part of the paper, the model of stratospheric influence on weather regimes is developed using ERA-Interim re-analysis data. As in previous studies, the limited observational record makes quantifying some parts of the stratospheric influence on weather regimes difficult and so in the second part of the paper we also make use of a large set of simulations from the ECMWF Integrated Forecasting System model to be able to further quantify this influence. We first show that the model faithfully reproduces the stratospheric influence on weather regimes determined from the re-analysis data. As shown by Dawson *et al.* (2012) it is important to use a relatively high resolution for this purpose, since climate resolution simulations are unable to faithfully reproduce the same weather regimes observed in re-analysis data. We use a model configuration (Ferranti *et al.* 2015) which has been previously shown to faithfully reproduce weather regime behaviour in the North Atlantic for this purpose.

## 2. Methodology

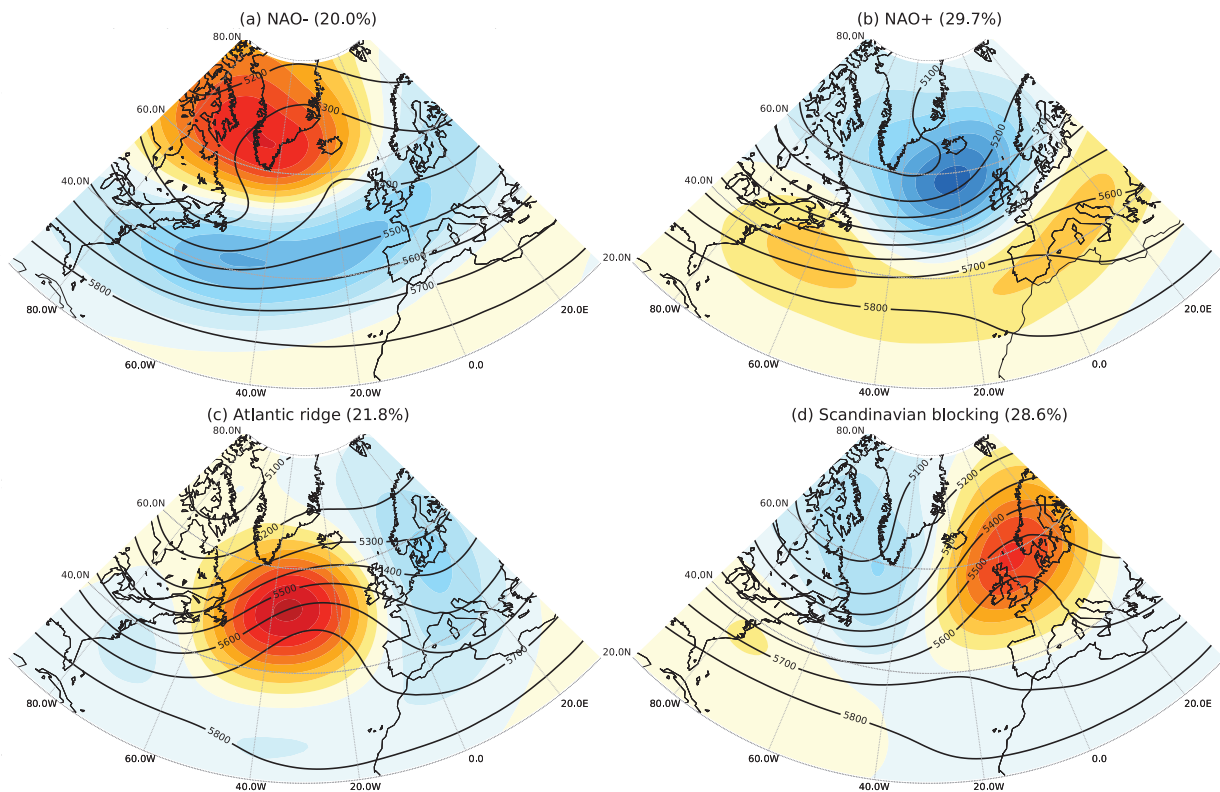
In all analysis in this study, data from the extended winter season (NDJFM) is considered.

### 2.1. Regime identification

The methodology used to derive weather regimes in both the ERA-Interim re-analysis data (Dee *et al.* 2011) and the IFS re-forecasts is similar to that used by Ferranti *et al.* (2015) and previously by Michelangeli *et al.* (1995) and Cassou (2008). Briefly, regimes are calculated as 'k-means' clusters of the phase space spanned by the first 14 leading Empirical Orthogonal Functions (EOFs) of the 500hPa geopotential height field over the North Atlantic domain (90°W-30°E; 20-80°N) for the cold season (November to March). For the re-analysis data, the winter seasons from 1979-1980 to 2016-2017 are used, giving a total of 38 winter seasons. The choice of 14 EOFs and 4 clusters is standard in the literature, but the details of the clustering are insensitive to the particular choices of these parameters (not shown).

87 This analysis produces the four familiar cluster patterns shown in Fig. 1 which are henceforth labelled as the Positive NAO (NAO+),  
 88 Negative NAO (NAO-), Scandinavian Blocking (BL) and Atlantic Ridge (AR) regimes. Three of the weather regimes describe the  
 89 three preferred North Atlantic jet stream locations (Woollings *et al.* 2010b), namely, NAO-, NAO+ and AR broadly correspond to  
 90 southern, central and northern jet-states respectively. Recent work also links the BL regime to a fourth mixed jet state and highlights  
 91 the complexity of comparing regimes from this kind of clustering approach and measures of the jet latitude (Madonna *et al.* 2017).  
 92 Other studies have emphasised the correlation between jet shifting and strengthening which can be observed when analysing the first  
 93 and second EOFs of the Atlantic zonal wind (Sparrow *et al.* 2009; Sheshadri and Plumb 2017).

94 The four regimes are used in the ECMWF medium-range clustering products (Ferranti and Corti 2011) to provide additional  
 95 information about the ensemble in terms of large-scale circulation and to allow an objective verification of the regime transitions.



**Figure 1.** Cluster centroids for the four regimes used in the study. Colours show 500hPa geopotential height anomalies for the centroid for the (a) NAO- (b) NAO+ (c) Atlantic Ridge and (d) Scandinavian Blocking regimes. For reference, the average geopotential height field is plotted for each regime in black contours. The percentage of days in which each regime occurs is shown at the top of each panel, this may not add up to 100% due to rounding

96 In both the re-analysis and model datasets each day is assigned to one of these four regimes based on the minimum Euclidian distance  
 97 between the regime centroid and the daily 500hPa geopotential height anomaly as described in Ferranti *et al.* (2015). In some cases, the  
 98 projection of an individual daily pattern onto all four of the centroids can be relatively weak and some authors have chosen to include  
 99 a separate 'no regime' classification in order to remove these cases from subsequent analysis (e.g. Grams *et al.* 2017). In our analysis  
 100 we choose not to take this approach, but do comment on the strength of the projection onto each cluster centroid for particular cases  
 101 throughout the manuscript.

102 The model dataset used is a set of ensemble reforecasts from the ECMWF Integrated Forecast System (IFS, cycle 40R1, operational  
 103 in 2014 with an enhanced ensemble size of 11 ensemble members). The data used covers 20 cold seasons from November 1994 to  
 104 February 2014. All reforecasts are made with the same version of the model, thereby avoiding any discontinuities related to changes in  
 105 model configuration. The first 14 days of each hindcast is discarded, so that the tropospheric evolution in each hindcast can be thought  
 106 of as an independent sample of the influence of the stratospheric state on weather regime frequency.

## 107 2.2. Stratospheric polar vortex strength

108 Throughout the manuscript we characterise the strength of the stratospheric polar vortex by calculating the daily-mean, zonal-mean,  
 109 zonal wind anomaly at 100hPa and 60° N. While other, more complex, measures might also be used to calculate the strength of the polar  
 110 vortex in the lower stratosphere, for example the annular mode index, the correlation between these measures and a simple diagnostic  
 111 of the zonal mean zonal wind is very high. In the first-order Markov models which are developed, the daily state of the polar vortex is  
 112 assigned to one of three tercile categories, where the thresholds for these categories are calculated from the full set of daily zonal mean  
 113 zonal wind anomaly in either the re-analysis data or model. For the ERA-Interim data, the threshold for the weak vortex state is  $-2.30$   
 114  $ms^{-1}$  and for the strong vortex state is  $2.19 ms^{-1}$ .

## 115 3. Results

### 116 3.1. Regime frequency changes in response to the stratosphere

117 As an initial exploration of the influence of stratospheric vortex strength on weather regimes in the North Atlantic, changes to the  
 118 frequency of occurrence, persistence and transition between regimes are first estimated. The frequency of weather regime occurrence

119  $(\hat{p})$  is a widely used metric (e.g. Cassou 2008) and is estimated here simply by dividing the number of times a given regime occurs (a)  
 120 by the total number of days sampled in each subset (n).

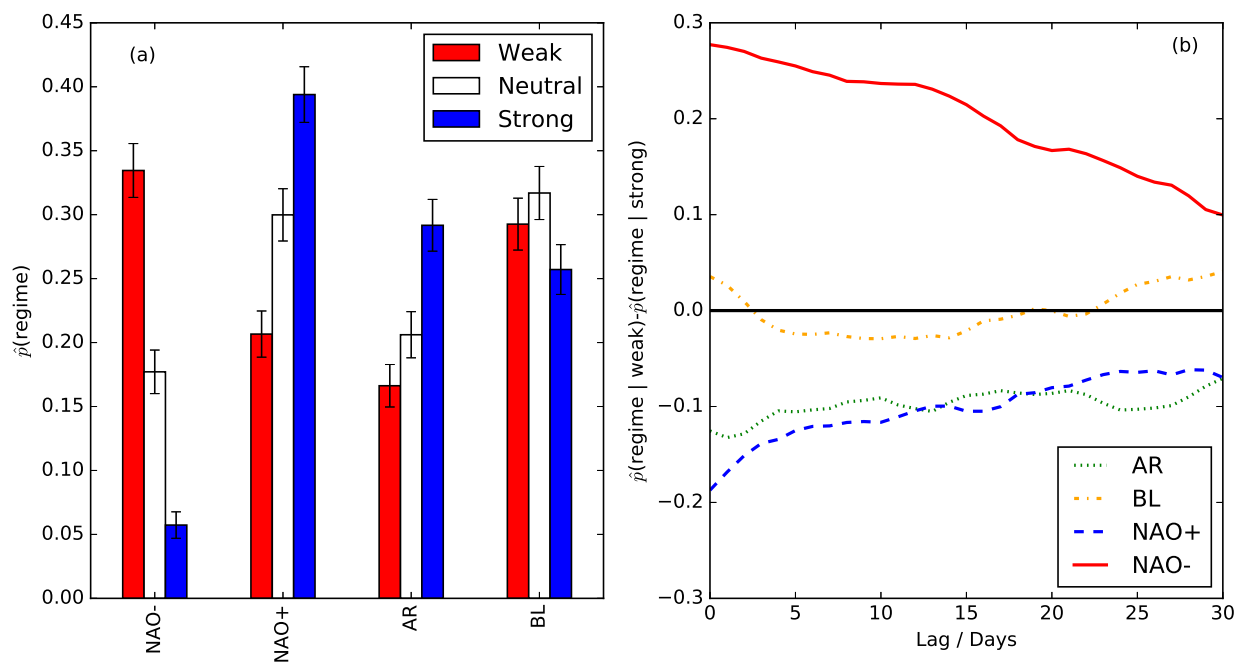
$$\hat{p} = \frac{a}{n} \quad (1)$$

121 A standard 95 % Wald confidence interval for the frequency is calculated since the sample size is large:

$$\hat{p} \pm 1.96 \sqrt{\frac{1}{n} \hat{p}(1 - \hat{p})} \quad (2)$$

122 To estimate the influence of the stratospheric vortex strength on the frequency of occurrence of weather regimes, this metric is  
 123 calculated for three sub-sets of the data based on the state of the polar vortex (weak, neutral and strong) on the previous day. The  
 124 frequency of occurrence estimates are shown in Fig. 2(a). This frequency is written as (where  $R_i$  represents one of the four weather  
 125 regimes):

$$\hat{p}(R_i(t+1)|U_{100}(t)). \quad (3)$$



**Figure 2.** (a) Frequency of occurrence of North Atlantic weather regimes immediately following weak stratospheric vortex conditions (left, red bar), neutral stratospheric vortex conditions (central, white bar) and strong stratospheric vortex conditions (blue, right bar). The 95% confidence interval (see text for details) is shown at the end of each bar. (b) Difference between frequency of occurrence of each regime in weak and strong stratospheric vortex conditions as a function of the lag between the stratospheric vortex conditions and the tropospheric regime. For example, a lag of 20 indicates that the frequency of occurrence of each tropospheric regime is conditional upon the stratospheric state twenty days before. Each line shows a different tropospheric regime, NAO- (red, solid), NAO+ (blue, dashed), AR (green, dotted) and BL (orange, dot-dash).

126 For neutral conditions, BL is the most frequently observed regime, followed by NAO+. Stratospheric vortex conditions significantly  
 127 shift weather regime frequency as expected. Following weak stratospheric vortex conditions, NAO- occurs almost twice as frequently  
 128 as during neutral stratospheric vortex conditions with consequent reductions in the occurrence frequency of both AR and NAO+.  
 129 Following strong stratospheric vortex conditions, the opposite sensitivity is found, large reductions in NAO- frequency and increases  
 130 in both AR and NAO+. Changes to BL frequency are smaller and not significant. Previous studies have highlighted the links between  
 131 Greenland Blocking and stratospheric vortex conditions (Woollings *et al.* 2010a; Davini *et al.* 2014), reflected here in the sensitivity of  
 132 the NAO- regime.

133 Since the lower stratosphere has a long memory during winter, the qualitative sensitivity of tropospheric weather regimes to  
 134 antecedent stratospheric conditions can be reproduced for a large range of lags between the variable chosen to represent the stratosphere  
 135 and the tropospheric weather regime. Fig. 2(b) shows the difference in occurrence frequency between weak and strong stratospheric  
 136 vortex states as a function of lag between the stratosphere and troposphere. A lag of twenty days in this figure indicates the estimated  
 137 frequency of occurrence of each tropospheric regime is conditional upon the stratospheric vortex state twenty days previously.  
 138 As expected, the difference between occurrence frequency of the NAO-, NAO+ and AR regimes declines as the lag between the  
 139 stratospheric vortex state and tropospheric weather regime increases. However, a large difference remains for the NAO-, NAO+ and  
 140 AR remains even at 30 days lag, consistent with the sensitivity to the stratospheric state shown in Fig. 2(a). For the remainder of the  
 141 manuscript we therefore continue to consider weather regime sensitivity to the stratospheric vortex state on the previous day.

142 The regime sensitivity to the antecedent stratospheric vortex state could be related either to changes to the probability of transition  
 143 to the AR, NAO+ and NAO- regimes or changes to probability of persistence of the same regimes. These quantities are estimated by  
 144 further sub-setting the observed regime data used in Fig. 2 adding a condition related to the weather regime on the previous day in  
 145 addition to the stratospheric vortex state. For example, the NAO- regime days in the weak vortex subset are further subset into two sets

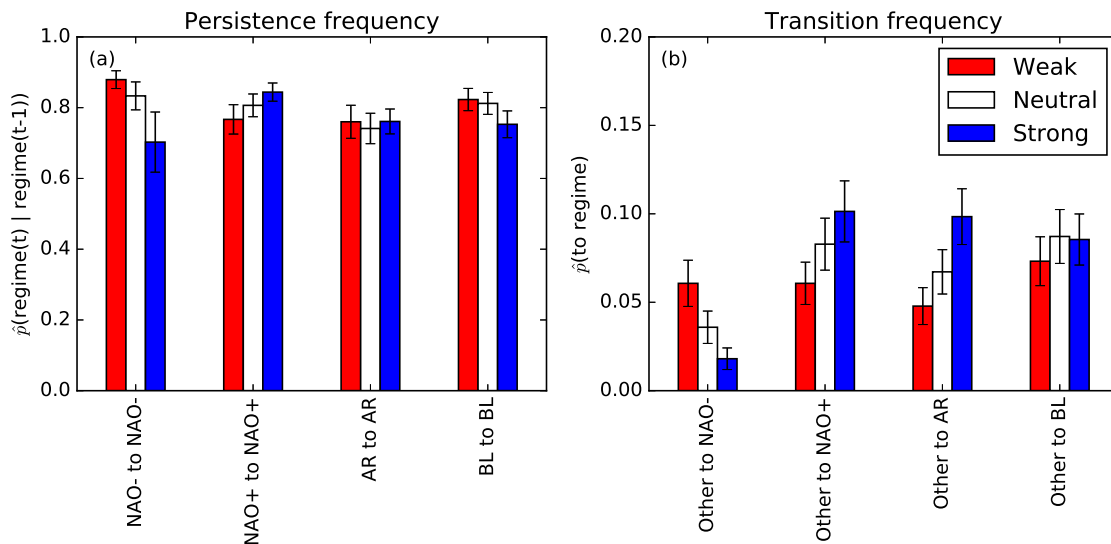


146 in which the weather regime on the previous day was NAO- and cases where the weather regime on the previous day was any other  
 147 regime. The number of days in which the previous regime was NAO- is used to estimate the persistence probability for the NAO-/weak  
 148 vortex case.

$$\hat{p}(R_i(t+1)|R_i(t) \& U_{100}(t)) \quad (4)$$

149 The number of days in which the previous regime was not NAO- is used to estimate the transition probability for the NAO-/weak  
 150 vortex case.

$$\hat{p}(R_i(t+1)|R_{j \neq i}(t) \& U_{100}(t)) \quad (5)$$



**Figure 3.** (a) Persistence and (b) transition frequency for weather regimes immediately following weak stratospheric vortex conditions (left, red bars), neutral stratospheric vortex conditions (central, white bars) and strong stratospheric vortex conditions (right, blue bars). The 95% confidence interval (see text for details) is shown at the end of each bar.

151 Figure 3 shows the persistence and transition probabilities for each of the four weather regimes following, weak, neutral and strong  
 152 stratospheric vortex conditions. The changes to persistence and transition probabilities are consistent with what might be expected  
 153 given the changes in regime frequency shown in Fig. 2, at least for the two NAO regimes. For antecedent weak vortex conditions,  
 154 the NAO- regime is significantly more persistent and a transition to this regime is more likely when compared to antecedent strong  
 155 vortex conditions. The NAO+ regime has an opposite sensitivity, with significantly enhanced persistence and transition frequency for  
 156 antecedent strong vortex conditions compared to weak vortex conditions. The AR regime has the same sensitivity as the NAO+ regime  
 157 in its transition frequency but no significant sensitivity in its persistence.

### 158 3.2. A simple Markov model

159 It is helpful to the subsequent discussion to cast the weather regime coupling to the stratosphere revealed in the previous section in  
 160 terms of a simple Markov model. An example for the NAO- regime is shown in Fig. 4. Fig. 4 shows the North Atlantic state as being  
 161 either NAO- or any one of the other regimes (other state in the figure). The arrows indicate persistence probability for NAO- ( $1-\alpha$ ) and  
 162 the transition probabilities between the NAO- state and other states ( $\alpha&\beta$ ), which can be estimated from the analysis in the previous  
 163 section.

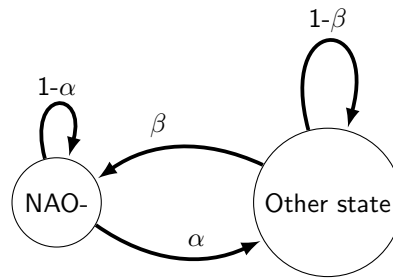
164 Describing North Atlantic regimes in this way assumes that state tomorrow depends only upon the state today. The model is shown  
 165 on the right hand side of Fig. 5 and is defined by the transition matrix, P, which in this case is:

$$P = \begin{bmatrix} 1 - \alpha & \alpha \\ \beta & 1 - \beta \end{bmatrix} = \begin{bmatrix} 0.83 & 0.17 \\ 0.04 & 0.96 \end{bmatrix} \quad (6)$$

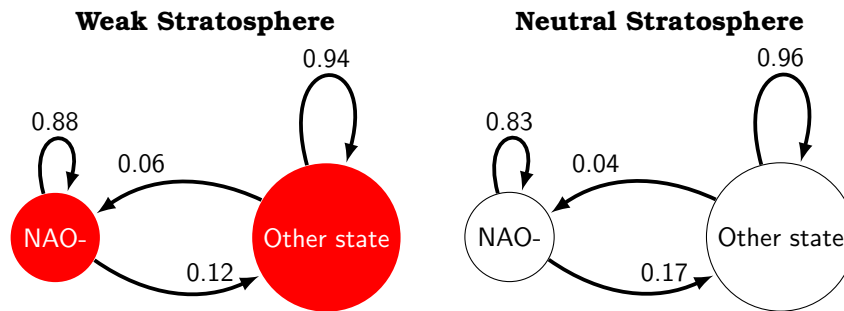
166 After N-steps of the model, the probability of being in either the NAO- or other state can be calculated from the N-th power of P.  
 167 In the limit of large n, the computed probability represents the steady state probability of the two states. For the case with a neutral  
 168 stratospheric vortex above:

$$\lim_{N \rightarrow \infty} P^N = \begin{bmatrix} \frac{\beta}{\alpha+\beta} & \frac{\alpha}{\alpha+\beta} \\ \frac{\beta}{\alpha+\beta} & \frac{\alpha}{\alpha+\beta} \end{bmatrix} = \begin{bmatrix} 0.19 & 0.81 \\ 0.19 & 0.81 \end{bmatrix} \quad (7)$$

169 The model described by P represents the observed frequency of occurrence of the NAO- state when the stratospheric vortex is neutral  
 170 very well ( $\hat{p} = 0.19$ ). To describe the impact of changes to the stratospheric state, an alternate Markov model that reflects the increased  
 171 persistence and likelihood of transition into the NAO- regime, is shown on the left hand side of Fig. 5.



**Figure 4.** Model of transitions between the NAO- and other states for the Neutral case. Circles indicate the weather regime. Arrows indicate transitions between states with the labels indicating the probability of following each path.



**Figure 5.** Model of transitions between the NAO- and other states for the weak and neutral stratospheric regimes. Circles indicate the weather regime. Arrows indicate transitions between states with the labels indicating the probability of following each path.

172 In this case, the steady state probability (for the weak vortex state) is:

$$\lim_{N \rightarrow \infty} P^N = \begin{bmatrix} 0.33 & 0.67 \\ 0.33 & 0.67 \end{bmatrix} \quad (8)$$

173 For a positive recurrent Markov chain like the one described above the mean recurrence time (the expected time taken to return to  
174 a given state) is given by the reciprocal of the steady state probability. For the two models discussed above, the recurrence time for  
175 the NAO- state is approximately five days when the stratospheric vortex is in a neutral state ( $\hat{p}=0.19$ ,  $1/\hat{p}=5.3$ ), three days when the  
176 stratospheric vortex is in a weak state ( $\hat{p}=0.33$ ,  $1/\hat{p}=3.0$ ) and sixteen days when the stratospheric vortex is in a strong state ( $\hat{p}=0.063$ ,  
177  $1/\hat{p}=15.9$ ).

178 The simple Markov chain model introduced here is an over-simplification of the real regime behaviour and it's connection to the  
179 stratosphere, but nonetheless helps to illustrate why diagnosing stratosphere-troposphere coupling is difficult with the limited data  
180 record available. Typically, daily composites following SSW or strong vortex events are used to diagnose the strength and extent of  
181 stratosphere-troposphere coupling in re-analysis datasets and model integrations. The analysis above suggests that for composites of  
182 30-100 events, daily estimates of the strength of the coupling will be strongly influenced by noise from the underlying tropospheric  
183 regime dynamics. A simple estimate of the size of this effect can be made numerically by simulating a large number of 60-day Markov  
184 chains with the same properties as the weak vortex chain above and constructing daily composites of the NAO- frequency.

185 For a large enough sample size, the daily average frequency of the NAO- regime should be 0.33 for each of the days in the composite.  
186 Variations in the daily NAO- frequency related to insufficient sampling can be quantified by calculating the daily standard deviation.  
187 In a toy experiment in which this process was repeated 100 times for a sample of 30 weak vortex events, the average daily standard  
188 deviation is 0.08. Even for a sample size of 100 events, the average daily standard deviation is 0.05. This result implies that composites  
189 of stratosphere-troposphere coupling events (e.g. [Baldwin and Dunkerton 2001](#)) likely contain significant sampling noise on the daily  
190 timescale and should only be used with some caution.

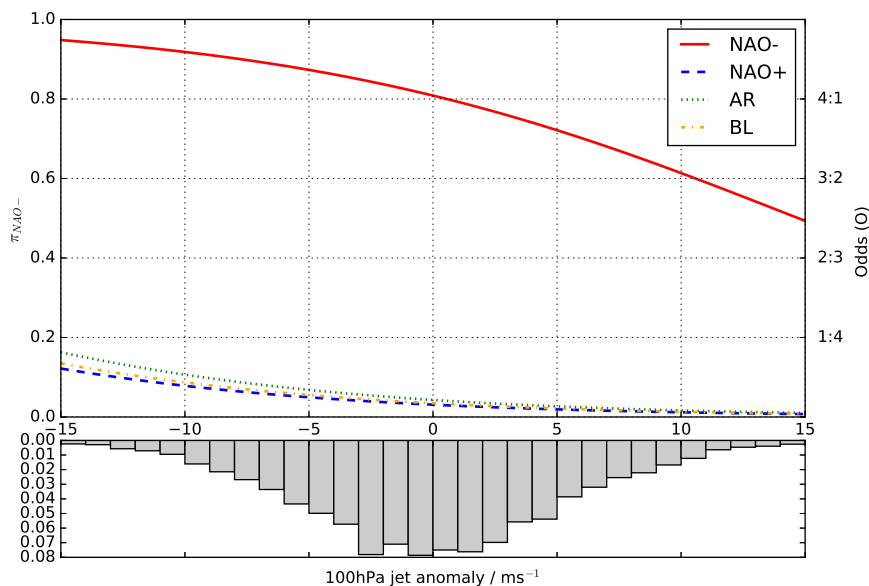
### 191 3.3. Quantifying stratosphere-troposphere coupling

192 In order to quantify coupling between the stratosphere and tropospheric weather regimes and compare coupling in different datasets  
193 and models it is necessary to develop a more comprehensive statistical model that can account for the effect of changes to the strength  
194 of the stratospheric vortex and the likelihood of transition between different weather regimes.

195 For this purpose, a logistic regression model is used. Logistic models are standard tools used in many branches of applied sciences  
196 where binomial observations are encountered ([Peng et al. 2002](#)). In the case of the regime data, this naturally forms a set of binomial  
197 variables ( $R_i(t)$ ) for each regime where  $R_i(t)=1$  indicates the weather regime on day  $t$  is regime  $i$ . Here the  $i$  refers to each of  
198 the different weather regimes (NAO-, NAO+, AR, BL) in turn. The logistic model predicts the probability ( $\pi_i$ ) that a given set of  
199 antecedent conditions today (the current weather regime and strength of the zonal mean zonal wind at 100hPa) will lead to a particular  
200 weather regime tomorrow. In order to deal with the non-linear nature of problems with binary response variables, a logistic regression  
201 model applies the logit transform to the dependent variable (in this case the probability of occurrence of a weather regime), where:

$$\text{logit}(O_i) = \ln \left( \frac{\pi_i}{1 - \pi_i} \right) \quad (9)$$





**Figure 6.** Model prediction for the NAO- regime for the logistic model fit to ERA-Interim data. Lines show the variation in the probability of occurrence of the NAO- regime as a function of the stratospheric wind anomaly. Solid, red line shows the probability when the previous state is also NAO-. The other coloured lines show the probability when the previous state is NAO+ (dashed, blue line), AR (dotted, green line) and BL (dash-dotted, orange line). A histogram of the stratospheric wind anomaly is shown below the x-axis.

202 The regression model is then very similar to a standard Ordinary Least Squares regression:

$$\text{logit}(O_i(t+1)) = \alpha U_{100}(t) + \sum_{j \neq i} \beta_j R_j(t) + c. \quad (10)$$

203 In the above expression,  $R_j(t)$  is a binary indicator variable for each of the weather regimes other than the one which the model  
 204 predicts. The model is fitted using each of the four weather regimes in turn. The parameters of the model ( $\alpha, \beta_j, c$ ) are then used  
 205 to diagnose the strength of stratosphere-troposphere coupling for each regime. Note that this model is linear in the log-odds of the  
 206 predicted regime but not the probability. (Typically, parameters of the model are expressed as an odds-ratio, which for the  $\beta_j$  parameters  
 207 is the ratio between the odds when the prior regime does and does not occur.

208 The nature of the data means that this model cannot simply be fit in the standard manner using a maximum likelihood estimate.  
 209 Since daily data for each winter season is used, it is strongly autocorrelated within each season (breaking the required assumption of  
 210 independence) but has little autocorrelation between seasons. Therefore, we make use of an alternative quasi-likelihood method, the  
 211 Generalized Estimating Equations (Ziegler and Vens 2010) which explicitly allows for this correlation structure when fitting the model.  
 212 The model is implemented and fit using the statsmodels (Seabold and Perktold 2010) and pandas libraries (McKinney 2010), freely  
 213 available for python.

214 An example of the diagnosis of coupling which can be gained from the model is shown in Fig. 6. The red line in this figure shows the  
 215 predicted probability of occurrence of the NAO- regime as a function of the stratospheric wind anomaly ( $\pi_{\text{NAO-}}$ ), when the existing  
 216 regime is also NAO-. As expected, the model shows a large increase in probability of the NAO- regime when stratospheric winds are  
 217 weak and a reduced probability when stratospheric winds are strong. The other coloured lines show the predicted occurrence probability  
 218 of the NAO- regime when the existing regime is NAO+, AR or BL. There are small differences between the three antecedent regimes,  
 219 but the largest difference is between regime persistence (the red line) and regime transition. For example, when stratospheric wind  
 220 anomalies are zero, the probability of transition from any of the three other regimes to NAO- is small (around 0.04). Therefore the  
 221 model captures a similar picture of the coupling as that shown in Fig. 4 and Fig. 5.

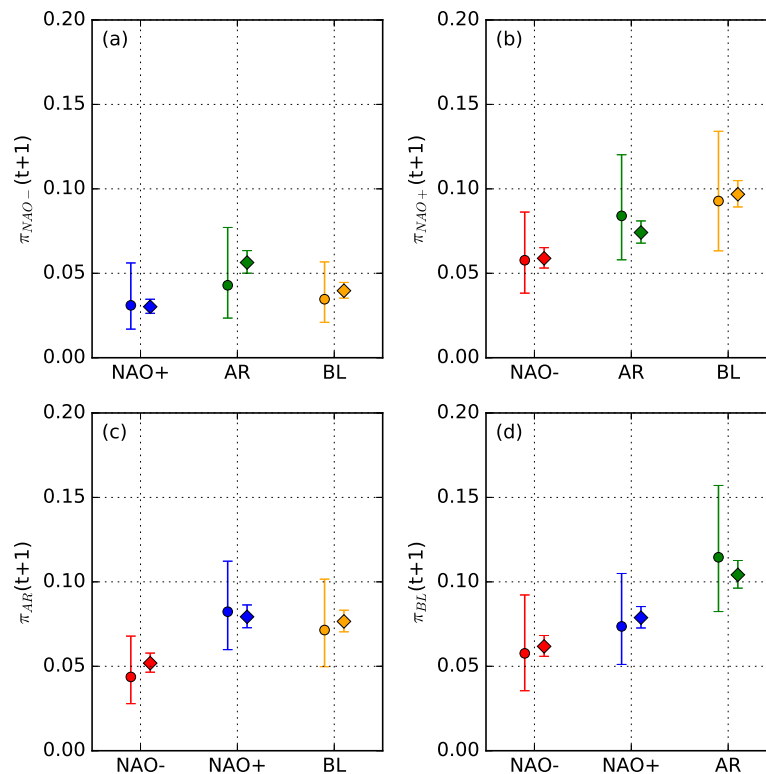
222 Comparison of the behaviour for all four regimes can be made by fitting the model with each of the regimes as the dependent  
 223 variable in turn. Central estimates for the model parameters ( $\alpha, \beta_j, c$ ) are shown in Table 1. Stratosphere-troposphere coupling  
 224 for each of the regimes is described by the odds ratio for the zonal wind predictor. For the NAO- case, the odds ratio  
 225 ( $O_{(\text{NAO-}|U_{100}=5)}/O_{(\text{NAO-}|U_{100}=0)}$ ) is 0.61 for a  $5\text{ms}^{-1}$  change in the stratospheric zonal wind anomaly (approximately one standard  
 226 deviation). Comparing with the red line in Fig. 6 helps to interpret the odds ratio. In Fig. 6 the probability of occurrence of the NAO-  
 227 state for a wind anomaly of  $0\text{ms}^{-1}$  and a prior NAO- state (red line) is approximately 0.8, or odds of 4:1. The diagnosed odds ratio  
 228 of 0.61 indicates that the odds of the NAO- state for a wind anomaly of  $5\text{ms}^{-1}$  should be 61% of 4:1 which are odds of 2.4:1 or a  
 229 probability of approximately 0.7 as shown by the red line.

230 The models for the NAO+ and AR regimes have an odds ratio greater than one for the stratospheric wind anomaly as expected,  
 231 indicating increased probability of occurrence for positive wind anomalies. However, the coupling strength for these regimes is much  
 232 smaller than for the NAO- regime, a 16% increase in the odds of NAO+ and AR regimes for a  $5\text{ms}^{-1}$  change in the stratospheric wind.  
 233 Furthermore, the BL regime does not show significant sensitivity to the stratospheric wind anomaly.

234 These results are consistent with the idealised modelling study of Garfinkel *et al.* (2013). The NAO- regime (typically associated  
 235 with a jet located around  $40^\circ\text{N}$  (Madonna *et al.* 2017) is more sensitive to stratospheric perturbations than regimes associated with jets  
 236 located further poleward (NAO+ and AR).

Table 1. Central estimates of model parameters for ERA-Interim data. Where an estimated parameter is not significant at a p-value of 0.05 it is written in *italic*. The intercept is shown as the probability of remaining in the dependent weather regime when the stratospheric wind anomaly is zero (it can be transformed back into the model parameter through the logit transform). The parameter for the stratospheric zonal mean zonal wind anomaly is shown as the odds ratio ( $O_{(i|U_{100}=5)}/O_{(i|U_{100}=0)}$ ) of the change for a  $5m.s^{-1}$  (an approximately 1 s.d.) change in wind anomaly. Parameters for weather regimes are shown as the odds ratio ( $O_{(i|j=j)}/O_{(i|j \neq j)}$ ). An odds ratio of one indicates no change in the odds for a given model predictor. Odds ratios greater than one indicate increases in the odds of the dependent weather regime. Odds ratios less than one indicate decreases in the odds of the dependent weather regime.

	NAO-(t+1)	NAO+(t+1)	AR(t+1)	BL(t+1)
Intercept	0.81	0.81	0.80	0.75
Stratospheric zonal wind(t)	0.61	1.16	1.16	<i>0.93</i>
NAO-(t)		0.01	0.02	0.02
NAO+(t)	0.01		0.03	0.02
AR(t)	0.01	0.02		0.03
BL(t)	0.01	0.02	0.03	



**Figure 7.** Model estimated transition probabilities for the (a) NAO-, (b) NAO+, (c) AR and (d) BL regimes from each of the other regimes. Filled dots show the estimate from the ERA-Interim dataset and diamonds show the estimate for the ECMWF IFS dataset. Whiskers on each estimate show the 95% confidence interval.

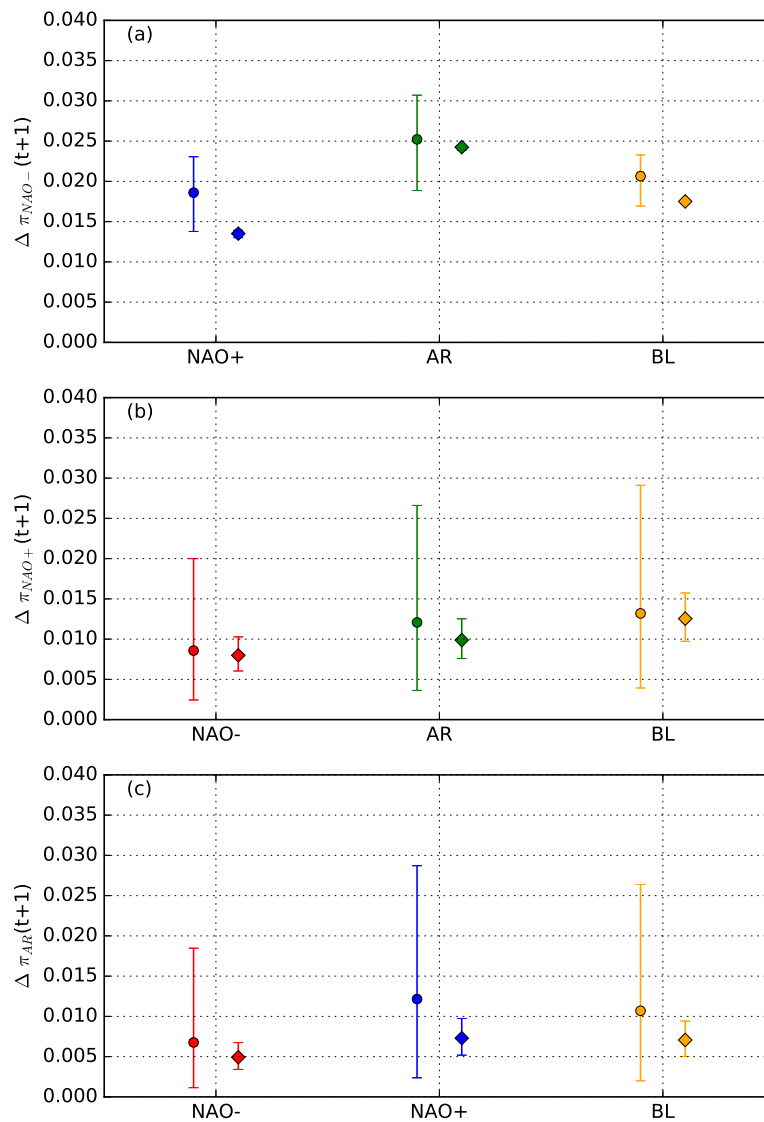
237 As the odds ratios for regime transitions are so small, reflecting the low-frequency memory of the weather regimes as described by  
 238 the simple Markov model, it is helpful to calculate and compare changes to the occurrence probability for a specific stratospheric state  
 239 ( $\pi_j$ ). Figure 7 shows model estimates of the transition probabilities for each of the four regimes. The transition probabilities are largest  
 240 for the NAO+ and BL regimes, reflecting the higher occurrence of these regimes in the dataset, and are smallest for the NAO- regime.  
 241 The transition probabilities are consistent with the right panel of Fig. 3 giving confidence in the fit of the model. For the ERA-Interim  
 242 data, although, for example in the NAO- case, the transition probability from AR to NAO- is larger than NAO+ to NAO-, the confidence  
 243 intervals for the two estimates overlap. When considering transitions to all four regimes, similar arguments hold, it appears that some  
 244 regime transitions are preferred above others but this cannot be conclusively stated given the limited re-analysis data record available.  
 245 We will return to this point when discussing the ECMWF IFS dataset below.

246 The model can also be used to diagnose how stratospheric wind anomalies influence transition probabilities for each regime. For  
 247 example for the NAO- regime, the model estimate of the impact of an approximately one standard deviation change in the stratospheric  
 248 wind anomaly can be calculated from:

$$\Delta p(\text{NAO} - (t+1)) = p(\text{NAO} - (t+1)|R_j(t) = 1 \& U_{100}(t) = -5) - p(\text{NAO} - (t+1)|R_j(t) = 1 \& U_{100}(t) = 0) \quad (11)$$

249 This quantity is shown in Fig. 8(a). Equivalent quantities for the NAO+ and AR regimes (although this time representing a difference  
 250 in probability between a positive  $5m.s^{-1}$  and zero lower stratospheric wind anomaly) are shown in Fig. 8(b) and (c).

251 The changes in transition probability associated with a one standard deviation zonal mean zonal wind anomaly for the NAO- regime  
 252 are much larger than for the NAO+ regime as expected from Table 1. The average sensitivity of the transition probability to stratospheric  
 253 wind anomalies is consistent with the simple analysis shown in Fig. 3. Changes of the transition probability are between 0.015 and  
 254 0.025 for transitions to the NAO- regime, a fractional change of 40-60% while changes of the transition probability for transitions to



**Figure 8.** Model estimated impact of an approximately one standard deviation ( $5\text{ms}^{-1}$ ) change to the stratospheric zonal wind anomaly on the transition probabilities for the (a)NAO-, (b) NAO+ and (c) AR regimes from each of the other regimes. Filled dots show the estimate from the ERA-Interim dataset and diamonds show the estimate for the ECMWF IFS dataset. Whiskers on each estimate show the 95% confidence interval. For the NAO- regime the estimated change is from 0 to  $-5\text{ms}^{-1}$  and for the NAO+ regime the estimated change is from 0 to  $+5\text{ms}^{-1}$  in order that the changes have the same sign.

255 the NAO+ regime are between 0.005 and 0.015, a fractional change of 10-30%. Very similar changes occur for transition to the AR  
 256 regime (Fig. 8(c))

### 257 3.4. Stratosphere-troposphere coupling in the ECMWF model

258 The same statistical model used to diagnose stratosphere-troposphere coupling in the ERA-Interim dataset can also be fit to the larger  
 259 ECMWF IFS reforecast dataset. The purpose of this analysis is two-fold, firstly it provides a means of comparing coupling in the  
 260 ECMWF model to the re-analysis in order to understand if it is consistent with that diagnosed in the re-analysis. Secondly, if the  
 261 coupling is broadly consistent, then the much larger ECMWF IFS dataset can reveal differences in the transition probabilities between  
 262 different regimes which cannot be consistently diagnosed in the smaller ERA-Interim dataset.

263 An identical logistic model is fit to the reforecast dataset, using the same Generalized Estimating Equations fitting method. Clusters  
 264 in the IFS dataset are the individual ensemble members of each forecast, we assume that there is autocorrelation between consecutive  
 265 days of each forecast but no correlation between each forecast member.

266 Table 2 shows central estimates of the model parameters, analogously to the parameters shown in 1. All model parameters are  
 267 consistent between the two datasets, with overlapping confidence intervals (not shown). This indicates that both the weather regime  
 268 behaviour and stratosphere-troposphere coupling in the IFS model is representative of the re-analysis and so can be used as a useful  
 269 proxy for the real atmosphere.

270 Model estimates of the transition probabilities ( $\pi_i$ ) in the IFS model are shown as diamonds in Fig. 7. Central estimates for the  
 271 transition probabilities are quite similar for the ERA-Interim and IFS datasets, with the same preference for particular regime transitions  
 272 for all four model fits. The much larger dataset available for the IFS model, means that confidence estimates for the transition  
 273 probabilities are much smaller, giving confidence that the differences in transition probabilities between regimes, while small, are  
 274 robust. For the NAO- regime (Fig. 7(a)), the transition probability from the AR regime is largest, approximately double the transition  
 275 probability from the NAO+ regime. For the NAO+ regime (Fig. 7(b)), there is a clear ordering with the largest transition probability

Table 2. Central estimates of model parameters for ECMWF IFS model data. Where an estimated parameter is not significant at a p-value of 0.05 it is written in italic. The intercept is shown as the probability of remaining in the dependent weather regime when the stratospheric wind anomaly is zero (it can be transformed back into the model parameter through the logit transform). The parameter for the stratospheric zonal mean zonal wind anomaly is shown as the odds ratio ( $O_{(i|U_{100}=5)}/O_{(i|U_{100}=0)}$ ) of the change for a  $5\text{ms}^{-1}$  (an approximately 1 s.d.) change in wind anomaly. Parameters for weather regimes are shown as the odds ratio ( $O_{(i|j=j)}/O_{(i|j\neq j)}$ ). An odds ratio of one indicates no change in the odds for a given model predictor. Odds ratios greater than one indicate increases in the odds of the dependent weather regime. Odds ratios less than one indicate decreases in the odds of the dependent weather regime.

	NAO-(t+1)	NAO+(t+1)	AR(t+1)	BL(t+1)
Intercept	0.81	0.81	0.78	0.76
Stratospheric zonal wind(t)	0.68	1.15	1.10	<i>1.00</i>
NAO-(t)		0.01	0.03	0.02
NAO+(t)	0.01		0.03	0.02
AR(t)	0.01	0.02		0.03
BL(t)	0.01	0.03	0.03	

276 from the BL regime, followed by the AR and then the NAO- regime. For the AR regime, (Fig. 7(c)) the transition probability from the  
 277 NAO- regime is smaller than the transition probability from the NAO+ or BL regimes. For the BL regime (Fig. 7(d)), there is a clear  
 278 ordering in transition probability with highest probability for the AR regime, followed by the NAO+ and NAO- regime.

279 Finally, it is useful to compare the sensitivity of regime transitions to the stratospheric state in the IFS model with the re-analysis.  
 280 The diamonds in Fig. 8 show the change in the transition probability due to a  $5\text{ms}^{-1}$  change in the strength of the stratospheric jet.  
 281 In all cases, the change in transition probability is consistent with the estimates from the ERA-Interim analysis, but at the lower end  
 282 of the confidence interval. For the NAO- regime, there is a clear separation between the change in transition probability for the three  
 283 possible precursor regimes. Transitions from AR to NAO- regime are most affected by stratospheric changes ( $\Delta p \approx 0.025$ ) followed  
 284 by transitions from the BL regime to NAO- and then transitions from the NAO+ to NAO- regime. For the NAO+ and AR regimes,  
 285 even with the larger IFS dataset, similar significant differences in the impact of stratospheric winds on transition probabilities are not  
 286 present.

#### 287 4. Discussion and Conclusions

288 In this study, the influence of stratospheric conditions on the North Atlantic troposphere is analysed by considering changes to weather  
 289 regime frequency and transition. A clear sensitivity of the NAO-, NAO+ and AR regimes to stratospheric conditions is demonstrated,  
 290 consistent with that shown by Beerli (2017). The simple model of regime frequency as a one-step Markov process which was developed  
 291 in section 3.1, makes it possible to demonstrate that this change in regime frequency is the result of changes in persistence probability  
 292 for the NAO-, NAO+ and AR regimes and the probability of transition into the same three regimes.

293 With this model in mind, a more complex logistic regression model can be fit to both the ERA-Interim re-analysis and the ECMWF  
 294 IFS model. An important primary conclusion of this analysis is that the coupling between the stratosphere and North Atlantic in the IFS  
 295 model is consistent with the re-analysis. By fitting a logistic regression model it is possible to be confident about this result, because  
 296 it makes use of the full dataset in each case producing model fit parameters which are highly significant. A standard approach to  
 297 comparing stratosphere-troposphere coupling in models by using composites of extreme events in the stratosphere would use a much  
 298 smaller fraction of the data and would be hampered by the kinds of sampling uncertainties discussed in section 3.2.

299 The logistic model shows that there is significantly non-linearity in the response of the troposphere to stratospheric anomalies,  
 300 consistent with Garfinkel *et al.* (2013). The NAO- regime is the most sensitive to changes in the stratospheric state, with smaller but  
 301 still significant sensitivity for the AR and NAO+ regimes. The BL regime is not sensitive to the stratospheric state in either the ERA-  
 302 Interim and ECMWF IFS datasets. Furthermore, for the transitions to the NAO- regime, the antecedent North Atlantic regime is an  
 303 important determinant of the North Atlantic response. For example, transitions between the AR and NAO- are enhanced to a greater  
 304 extent than transitions between the NAO+ and NAO- for the same change in stratospheric winds. These two results are important both  
 305 for forecasters seeking to make use of stratosphere-troposphere coupling when diagnosing model responses for individual SSW or  
 306 strong vortex events and for model developers.

307 Understanding the dynamical reasons why the AR regime is more sensitive to stratospheric changes is beyond the scope of this  
 308 study, but recent work (e.g. Michel and Rivière 2011; Swenson and Straus 2017) which links regime transitions to changes in Rossby  
 309 Wave Breaking (RWB) points to a possible direction for further study. In particular, transitions to NAO- states are often preceded  
 310 by enhancement of cyclonic RWB over North America and Greenland, accompanied by weak zonal wind anomalies and enhanced  
 311 horizontal wind shear on the poleward side of the jet. Similar easterly zonal wind anomalies would be produced in the lower stratosphere  
 312 and upper stratosphere on the poleward side of the jet during SSW events.

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#### 318 References

319 Baldwin MP, Dunkerton TJ. 1999. Propagation of the Arctic Oscillation from the stratosphere to the troposphere. *Journal of Geophysical Research: Atmospheres*  
 320 **104**(D24): 30 937–30 946.

- 321 Baldwin MP, Dunkerton TJ. 2001. Stratospheric harbingers of anomalous weather regimes. *Science* **294**(5542): 581–584.
- 322 Beerli R. 2017. Sources of sub-seasonal predictability for energy industry-relevant weather events. *qjrms* (ETH No. 24488).
- 323 Beerli R, Wernli H, Grams CM. 2017. Does the lower stratosphere provide predictability for month-ahead wind electricity generation in Europe? *Quarterly*  
324 *Journal of the Royal Meteorological Society* **In press**.
- 325 Cassou C. 2008. Intraseasonal interaction between the Madden-Julian Oscillation and the North Atlantic Oscillation. *Nature* **455**(7212): 523.
- 326 Cassou C, Terray L, Hurrell JW, Deser C. 2004. North Atlantic winter climate regimes: Spatial asymmetry, stationarity with time, and oceanic forcing. *Journal*  
327 *of Climate* **17**(5): 1055–1068.
- 328 Davini P, Cagnazzo C, Anstey J. 2014. A blocking view of the stratosphere-troposphere coupling. *Journal of Geophysical Research: Atmospheres* **119**(19).
- 329 Dawson A, Palmer T, Corti S. 2012. Simulating regime structures in weather and climate prediction models. *Geophysical Research Letters* **39**(21).
- 330 Dee DP, Uppala S, Simmons A, Berrisford P, Poli P, Kobayashi S, Andrae U, Balmaseda M, Balsamo G, Bauer P, Bechtold P, Belijaars A, van de Berg L,  
331 Bidlot J, Bormann N, Delsol C, Dragani R, Fuentes M, Geer A, Haimberger L, Healey S, Hersbach H, Holm E, Isaksen L, Kallberg P, Kohler M, Matricardi  
332 M, McNally A, Monge-Sanz B, Morcrette JJ, Park BK, Peubey C, de Rosnay P, Tavolato C, Thepaut JN, Vitart F. 2011. The ERA-Interim reanalysis:  
333 Configuration and performance of the data assimilation system. *Quarterly Journal of the Royal Meteorological Society* **137**(656): 553–597.
- 334 Ferranti L, Corti S. 2011. New clustering products. *ECMWF Newsletter* **127**: 6–11.
- 335 Ferranti L, Corti S, Janousek M. 2015. Flow-dependent verification of the ECMWF ensemble over the Euro-Atlantic sector. *Quarterly Journal of the Royal*  
336 *Meteorological Society* **141**(688): 916–924.
- 337 Garfinkel CI, Waugh DW, Gerber EP. 2013. The effect of tropospheric jet latitude on coupling between the stratospheric polar vortex and the troposphere.  
338 *Journal of Climate* **26**(6): 2077–2095.
- 339 Grams CM, Beerli R, Pfenninger S, Staffell I, Wernli H. 2017. Balancing europe's wind-power output through spatial deployment informed by weather regimes.  
340 *Nature climate change* **7**(8): 557.
- 341 Hannachi A, Straus DM, Franke CL, Corti S, Woollings T. 2017. Low Frequency Nonlinearity and Regime Behavior in the Northern Hemisphere Extra-Tropical  
342 Atmosphere. *Reviews of Geophysics* .
- 343 Hitchcock P, Simpson IR. 2014. The downward influence of stratospheric sudden warmings. *Journal of the Atmospheric Sciences* **71**(10): 3856–3876.
- 344 Hitchcock P, Simpson IR. 2016. Quantifying Eddy Feedbacks and Forcings in the Tropospheric Response to Stratospheric Sudden Warmings. *Journal of the*  
345 *Atmospheric Sciences* **73**(9): 3641–3657.
- 346 Karpechko AY. 2015. Improvements in statistical forecasts of monthly and two-monthly surface air temperatures using a stratospheric predictor. *Quarterly*  
347 *Journal of the Royal Meteorological Society* **141**(691): 2444–2456.
- 348 Kidston J, Scaife AA, Hardiman SC, Mitchell DM, Butchart N, Baldwin MP, Gray LJ. 2015. Stratospheric influence on tropospheric jet streams, storm tracks  
349 and surface weather. *Nature Geoscience* **8**(6): 433.
- 350 Limpasuvan V, Hartmann DL, Thompson DW, Jeev K, Yung YL. 2005. Stratosphere-troposphere evolution during polar vortex intensification. *Journal of*  
351 *Geophysical Research: Atmospheres* **110**(D24).
- 352 Limpasuvan V, Thompson DW, Hartmann DL. 2004. The life cycle of the northern hemisphere sudden stratospheric warmings. *Journal of Climate* **17**(13):  
353 2584–2596.
- 354 Madonna E, Li C, Grams CM, Woollings T. 2017. The link between eddy-driven jet variability and weather regimes in the North Atlantic-European sector.  
355 *Quarterly Journal of the Royal Meteorological Society* .
- 356 McKinney W. 2010. Data structures for statistical computing in python. In: *Proceedings of the 9th Python in Science Conference*, van der Walt S, Millman J  
357 (eds), pp. 51 – 56.
- 358 Michel C, Rivière G. 2011. The link between rossby wave breakings and weather regime transitions. *Journal of the Atmospheric Sciences* **68**(8): 1730–1748.
- 359 Michelangeli PA, Vautard R, Legras B. 1995. Weather regimes: Recurrence and quasi stationarity. *Journal of the Atmospheric Sciences* **52**(8): 1237–1256.
- 360 Peng CYJ, Lee KL, Ingersoll GM. 2002. An introduction to logistic regression analysis and reporting. *The Journal of Educational Research* **96**(1): 3–14.
- 361 Scaife A, Karpechko AY, Baldwin M, Brookshaw A, Butler A, Eade R, Gordon M, MacLachlan C, Martin N, Dunstone N. 2016. Seasonal winter forecasts and  
362 the stratosphere. *Atmospheric Science Letters* **17**(1): 51–56.
- 363 Seabold J, Perktold J. 2010. Statsmodels: Econometric and statistical modeling with python. In: *Proceedings of the 9th Python in Science Conference*.
- 364 Sheshadri A, Plumb RA. 2017. Propagating annular modes: Empirical orthogonal functions, principal oscillation patterns, and time scales. *Journal of the*  
365 *Atmospheric Sciences* **74**(5): 1345–1361.
- 366 Sigmond M, Scinocca J, Kharin V, Shepherd T. 2013. Enhanced seasonal forecast skill following stratospheric sudden warmings. *Nature Geoscience* **6**(2):  
367 98–102.
- 368 Sparrow S, Blackburn M, Haigh JD. 2009. Annular variability and eddy-zonal flow interactions in a simplified atmospheric gcm. part i: Characterization of  
369 high- and low-frequency behavior. *Journal of the Atmospheric Sciences* **66**(10): 3075–3094.
- 370 Swenson ET, Straus DM. 2017. Rossby Wave Breaking and Transient Eddy Forcing during Euro-Atlantic Circulation Regimes. *Journal of the Atmospheric*  
371 *Sciences* **74**(6): 1735–1755.
- 372 Tripathi OP, Baldwin M, Charlton-Perez A, Charron M, Eckermann SD, Gerber E, Harrison RG, Jackson DR, Kim BM, Kuroda Y, Lang A, Mahmood S, Mizuta  
373 R, Roff G, Sigmond M, Son SW. 2015a. The predictability of the extratropical stratosphere on monthly time-scales and its impact on the skill of tropospheric  
374 forecasts. *Quarterly Journal of the Royal Meteorological Society* **141**(689): 987–1003.
- 375 Tripathi OP, Charlton-Perez A, Sigmond M, Vitart F. 2015b. Enhanced long-range forecast skill in boreal winter following stratospheric strong vortex conditions.  
376 *Environmental Research Letters* **10**(10): 104007.
- 377 Woollings T, Charlton-Perez A, Ineson S, Marshall A, Masato G. 2010a. Associations between stratospheric variability and tropospheric blocking. *Journal of*  
378 *Geophysical Research: Atmospheres* **115**(D6).
- 379 Woollings T, Hannachi A, Hoskins B. 2010b. Variability of the North Atlantic eddy-driven jet stream. *Quarterly Journal of the Royal Meteorological Society*  
380 **136**(649): 856–868.
- 381 Ziegler A, Vens M. 2010. Generalized estimating equations. *Methods of Information in Medicine* **49**(5): 421–425.
- 382 Zubiate L, McDermott F, Sweeney C, O'Malley M. 2017. Spatial variability in winter nao?wind speed relationships in western europe linked to concomitant  
383 states of the east atlantic and scandinavian patterns. *Quarterly Journal of the Royal Meteorological Society* **143**(702): 552–562, doi:10.1002/qj.2943, URL  
384 <http://dx.doi.org/10.1002/qj.2943>.