https://doi.org/10.5194/wcd-2022-39
Preprint. Discussion started: 21 July 2022

Weather and Climate Dynamics Discussions
Open Access

Using large ensembles to quantify the impact of sudden stratospheric warmings on the North Atlantic Oscillation

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Abstract. Sudden stratospheric warming events (SSWs) are often followed by significant weather and climate impacts at the surface. By affecting the North Atlantic Oscillation (NAO), SSWs can lead to periods of extreme cold in parts of Europe and North America. Previous studies have used observations and free-running climate models to try to identify features of the atmosphere prior to an SSW that can determine the subsequent impact at the surface. However, the limited observational record makes it difficult to accurately quantify these relationships. Here, we instead use a large ensemble of seasonal hindcasts. We first test whether the hindcasts reproduce the observed characteristics of SSWs and their surface signature. We find that the simulations are statistically indistinguishable from the observations, in terms of the overall risk of an SSW per winter (56%), the frequency of SSWs with negative NAO responses (65%), the magnitude of the NAO responses, and the frequency of wavenumber-2 dominated SSWs (26%). We also assess the relationships between prior conditions and the NAO response following an SSW. We find that there is little information in the precursor state to guide differences in the subsequent NAO behaviour between one SSW and another, reflecting the substantial natural variability between SSW events. The strongest relationships with the NAO response are from pre-SSW sea level pressure anomalies over the polar cap, and from zonal wind anomalies in the lower stratosphere, both exhibiting correlations of around 0.3. The pre-SSW NAO has little bearing on its post-SSW state. The strength of the pre-SSW zonal wind anomalies at 10 hPa is also not significantly correlated with the NAO response. Finally, we find that there is no significant difference in the likelihood of a post-SSW negative NAO response between wave-1 and wave-2 dominated SSWs, although the latter result in a stronger negative NAO anomaly on average.

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1 Introduction

Since their discovery (Scherhag, 1952), sudden stratospheric warmings (SSWs) have been recognised as some of the most dramatic events in the Earth's atmosphere. The strong winds of the winter stratospheric polar vortex are disrupted, and even reversed, by the breaking of planetary-scale Rossby waves propagating upwards from the troposphere. The resulting descent of air over the pole causes adiabatic warming, with temperatures rising by several tens of kelvin over a matter of days (e.g. Baldwin et al., 2021 and references therein). The zonal mean signature of this disruption propagates downwards (Kodera, 1995), leading in some cases to impacts at the surface (Baldwin and Dunkerton, 1999, 2001; Christiansen, 2001), enhanced by eddy feedbacks in a way that is only partly understood (Kidston et al., 2015; Kunz and Greatbatch, 2013). In the Northern Hemisphere, the changes in the tropospheric circulation are often characterised in terms of the Arctic Oscillation (AO) and the North Atlantic Oscillation (NAO), reflecting an equatorward shift of the jet stream and storm track (Baldwin and Dunkerton, 2001). This leads to corresponding impacts on the weather (e.g. Butler et al., 2017; King et al., 2019), with warm anomalies in eastern Canada, and cold anomalies in the eastern United States and across northern Eurasia. The northern European regions of Scandinavia and the British Isles experience reduced precipitation, while central and southern Europe experience wetter than average conditions. In East Asia, SSWs can affect the East Asian Winter Monsoon (Deng et al., 2008): Although China on average tends to experience milder conditions following an SSW (Lim et al., 2019), East Asia in general can see an increased risk of extreme cold air outbreaks (Huang et al., 2021; Kolstad et al., 2010; Song et al., 2015). In general, SSWs are linked to extremes in surface climate (e.g. Domeisen and Butler, 2020; Huang et al., 2021), with potentially severe impacts on human health and wellbeing (e.g. Charlton-Perez et al., 2021).

Although individual SSWs themselves are predictable in a deterministic sense on timescales of one to two weeks (Taguchi, 2016), their prolonged disruption of the stratosphere and impact at the surface means that the occurrence of an SSW can increase the predictability of the subsequent surface climate to one or two months (Scaife et al., 2022; Sigmond et al., 2013). The predictability of SSW events is improved in models with a better-resolved stratosphere (Marshall and Scaife, 2010), although this is a necessary but not sufficient condition for good representation of SSWs: Chávez et al. (2022) for example showed that using a coupled ocean model had more impact than vertical resolution. The overall winter risk of an SSW occurring can also be predicted probabilistically with some skill at lead times of several months (Scaife et al., 2016), although the skill is dependent on other concurrent climate features such as El Niño.

SSWs have been observed to occur approximately 6 times per decade in the Northern Hemisphere (Bancalá et al., 2012; Charlton and Polvani, 2007), and only about 2/3 of SSWs are followed by effects at the surface as described above (e.g. White et al., 2019). A number of studies have used observation-based data (reanalyses) to investigate precursors of SSWs in surface climate features, as well as in the wave driving from the troposphere and in characteristics of the vortex itself (e.g. Bao et al., 2017; Cohen and Jones, 2011; Domeisen et al., 2020; Martius et al., 2009; Mitchell et al., 2013; Nakagawa and Yamazaki, 2006; Polvani and Waugh, 2004; Seviour et al., 2013; Shen et al., 2020). However, the variability seen between different

SSWs, and in the climate conditions in which they occur, coupled with their relatively low frequency and the limited observational record (only a handful of decades), has made it difficult to make definitive statements on what, if any, preconditioning causes stronger surface impacts following one SSW than another.

Recognising the limited observational sample, some studies have sought to increase their sample sizes by using free-running climate models (e.g. Garfinkel et al., 2010; Karpechko et al., 2017; Kolstad et al., 2010; Kolstad and Charlton-Perez, 2011; Maycock and Hitchcock, 2015), performing dedicated model experiments (e.g. de la Cámara et al., 2017), or using "ensembles of opportunity" from climate model experiments designed for other studies (e.g. White et al., 2019). Although helpful, these approaches are not without problems. The climate model runs used might not be designed to simulate the climate over the same period as the observations, and might be subject to biases or trends that grow over the course of the runs. Different models will be subject to different biases, and this can make it difficult to interpret results in terms of uncertainty. Hall et al. (2022) demonstrated that although models from the recent Coupled Model Intercomparison Project 6 (CMIP6) ensemble performed well in terms of their responses to SSWs, they exhibited different tropospheric precursors compared to observations. Tyrrell et al. (2022) showed that the weak SPV in their model experiments resulted in too many SSWs, and showed that a nudging bias correction method could improve this. However, the PMSL response to SSWs in their model was neither biased, nor affected by their bias correction. An alternative approach is to bootstrap the observational sample itself (e.g. Oehrlein et al., 2021), which allows the uncertainty in the observed sample to be estimated. Nevertheless, the results do not necessarily span the full range of possible present-day climate variability due to the inherent limitations of the observed sample size.

As a result of these limitations, existing studies do not always agree. For example, Mitchell et al. (2013) and Seviour et al. (2013) found that whether an SSW is characterised by the vortex splitting or simply being displaced has a significant impact on the subsequent surface response. In contrast, Charlton and Polvani (2007), Cohen and Jones (2011), and Maycock and Hitchcock (2015) found that the differences were small, subject to sampling variability, and not robust to changes in methodology or data. The uncertainties brought about by the limited observational sample and compounded by possible errors in climate models have meant that there is still no real consensus on these questions.

Instead, in this study, we use a large ensemble of initialised climate simulations, produced as seasonal hindcasts. In contrast to free-running climate models, these initialised simulations are more closely constrained to describe variability within the recent observed climate, while also providing a much larger data set. This follows the UNSEEN approach (UNprecedented Simulated Extremes using ENsembles, Thompson et al., 2017), in which a large ensemble of initialised hindcasts is used to greatly increase sample sizes, to quantify the probability of plausible but unobserved climate states. In our case, we are not quantifying rare weather extremes, but are nevertheless interested in the likelihood of particular climate events and their responses. The model and observational data we use are detailed in Sect. 2, together with descriptions of how we characterise SSWs and their responses. In Sect. 3 we demonstrate the accuracy of our model data in representing SSWs and their surface impacts, in

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comparison to the observations and their sampling uncertainty. Sect. 4 examines what determines the NAO response to SSWs, by considering precursors at the surface and in the stratosphere. We discuss and summarise our results in Sect. 5.

2 Data and methods

2.1 Climate model hindcast and observation-based data

We use hindcast data from the GloSea5 seasonal forecast system (MacLachlan et al., 2015), which is based on the HadGEM3-GC2 coupled climate model (Williams et al., 2015). The model has an atmospheric grid spacing of 0.833° longitude and 0.556° latitude, and 85 vertical levels extending to a height of 85 km. GloSea5 has been shown to have a good representation of the total variance of the NAO (Scaife et al., 2014). The hindcasts cover 23 winters, 1993/94 to 2015/16. Daily data for an extended winter period (December–March, DJFM) are used from predictions initialised on three dates centred on early November (25th Oct, 1st Nov, 9th Nov). We have 14 ensemble members available per initialisation date. This therefore yields an ensemble of $14 \times 3 = 42$ members per winter, leading to an overall sample of $42 \times 23 = 966$ winters altogether. The observational record we use is from ERA-Interim (Dee et al., 2011), covering 40 winters from 1979/80 to 2018/19. This has a grid spacing of $0.75° \times 0.75°$, and 37 vertical levels.

In order to make fair comparisons between the model data and ERA-Interim, we resample the model data into a series of 1000 ensembles, each with the same number of winters as the observations. We can then test whether the single observational sample of 40 years is consistent with the distribution of possible 40-year samples seen in the model hindcasts, allowing us to account for sampling uncertainty due to climate variability. In each of the 1000 resamples, an ensemble member is randomly picked from the hindcast either from the same year (for the 23 years within the hindcast period), or from across the whole sample of 966 winters (for the remaining 17 years).

2.2 Methods

We use daily mean pressure at mean sea level (PMSL), zonal wind, and geopotential height (GPH) data. To calculate anomalies, we create daily climatologies for each variable, smoothed using a Gaussian filter with a standard deviation of 10 days.

We define an SSW event simply as the first time in a DJF period (ignoring any leap days) when the zonal mean zonal wind at 60° N and 10 hPa goes below zero (following Charlton and Polvani, 2007). We therefore measure zero or one SSW per winter. Defining SSWs only in DJF, rather than DJFM, helps separate genuine SSW events from the final warming at the end of a winter, as well as ensuring that we always have at least 30 days of data after every SSW to assess their subsequent impact. For some analyses, we will also require a number of days before each SSW, to assess the impact of precursor climate features. Typically this will be 10 days, which restricts our sample of SSWs to those occurring on or after 11th December. The 30-day

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post-SSW period and 10-day pre-SSW periods were chosen after examining the composite mean time series of data before and after the SSW events, discussed in the results below, although other time periods were also tested.

We focus on the impact of SSWs on the North Atlantic Oscillation (NAO). We define an NAO anomaly index as the difference between PMSL anomalies in two boxes, one near the Azores (28° W to 20° W, 36° N to 40° N) minus one over Iceland (25° W to 16° W, 63° N to 70° N), following Dunstone et al. (2016). In addition to the NAO anomalies before and after SSWs, we also consider the "climatological" NAO anomalies, i.e. independent of whether or not there is an ongoing SSW. For this we calculate the mean NAO anomalies in 5,000 random 30d periods that start in DJF (i.e. mirroring our SSW definition), from across our 966 model winters. The proportion of negative NAO anomalies can then be calculated (43%).

We characterise our SSW events according to the dominant zonal wavenumber in the vortex (cf. Martius et al., 2009; Nakagawa and Yamazaki, 2006). We calculate the amplitudes of zonal waves 1 and 2 (A_1 and A_2 respectively), from the Fourier transform of the daily mean eddy geopotential height at 60° N and 50 hPa (results using 10 hPa are similar). We focus on zonal wavenumbers 1 and 2, as there is very little contribution from higher wavenumbers at these altitudes, and consider SSW events with $A_2 > A_1$ on the date of the SSW to be wave-2 dominated, and those with $A_2 < A_1$ to be wave-1 dominated (results using the mean amplitudes over the 10 days pre-SSW are similar). SSWs dominated by wave 1 will tend to correspond more to "displacement" events, and wave-2 dominated events will involve a split in the vortex. However, all events will involve a mixture of different wavenumbers, and there is no direct correspondence between our wave-1/wave-2 dominated classification and a displacement/split classification based on vortex geometry (Martius et al., 2009; Seviour et al., 2013; White et al., 2019).

Finally, we calculate 95% confidence intervals on correlations using a Fisher Z transformation (e.g. Wilks, 2020), and on proportions/frequencies using a Wilson interval (e.g. Brown et al., 2001). We use a standard binomial test to compare a sample proportion to a binomial distribution, the standard Gaussian approach for binomial statistics to test if two sample proportions are significantly different to each other, and the standard student's *t* test to assess if two means are significantly different. These tests are all performed at the 5% level.

3 Does the seasonal forecast system represent stratosphere–troposphere coupling accurately?

Histograms showing the frequency of SSWs and their NAO responses are shown in Figure 1, with the overall SSW frequency shown in Figure 1(a). In the model, 545 winters out of 966 have at least one SSW, i.e. 56% (with a 95% confidence interval of 53%–60%, and significantly different to 50%). This compares with 19 winters out of 40 in ERA-Interim (48%, not significantly different to 50%). The 95% range from our model resamples is 41%–72%, covering the observed value, and giving an indication of its larger uncertainty due to the more limited sample size. The observed frequency is therefore statistically indistinguishable from the more robust estimate from the model hindcast.

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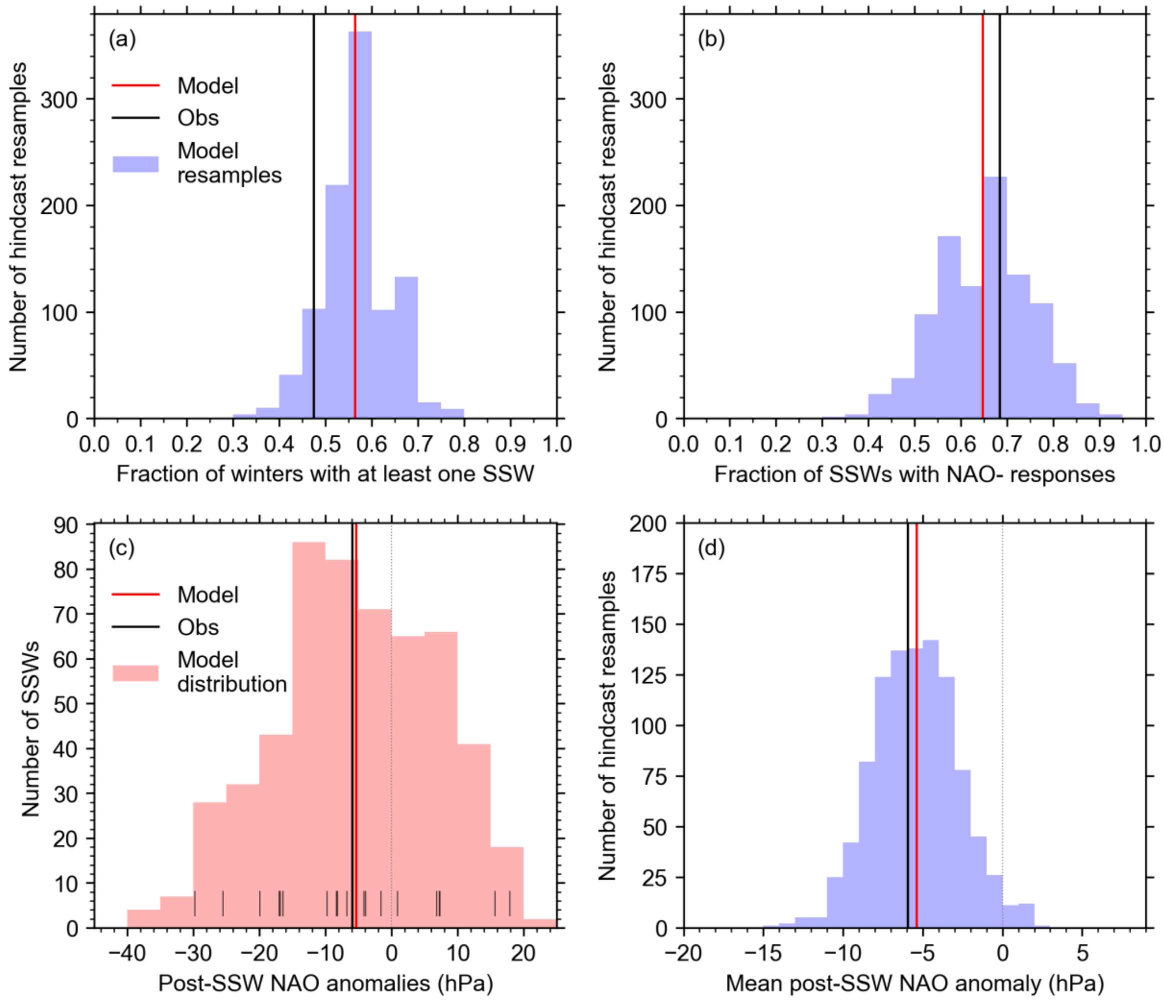

Figure 1. The frequency of SSWs and their NAO responses. Panels show (a) the fraction of winters with at least one SSW in DJF; (b) the fraction of those SSWs with negative NAO responses, based on the mean NAO anomaly in the 30d following the SSW; (c) the distribution of those post-SSW NAO anomalies, over all SSWs; and (d) the mean of those post-SSW anomalies, over all SSWs. All panels use the same colouring: Vertical lines show the single values from the observations (black) and the model (red), and blue histograms show the distributions of these over the model resamples. In panel (c) the red histogram shows the distribution over SSWs in the model, and the black ticks at the bottom of the plot show the observed distribution; the solid vertical lines in panels (c) and (d) show the same means, and the faint dotted vertical lines indicate zero.

The NAO responses are examined in terms of the mean NAO anomaly in the 30 days following an SSW. The frequency of a negative NAO response is shown in Figure 1(b) to be about 2/3, consistent with previous results (e.g. Jucker, 2016; Karpechko et al., 2017; Runde et al., 2016; White et al., 2019; although these vary widely depending on the methods and data used). The model has 353 SSWs that are followed by a 30d mean negative NAO, out of 545 (65%, with 95% confidence interval of 61%–69%). The observed proportion is very similar, at 13 SSWs out of 19 (68%). However, the model resamples suggest a much

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wider range of NAO-negative frequencies was possible from a 40-year sample like the observations, with a central (95%) range of 43% to 84%.

These frequencies can also be compared to the probability of 30d negative NAO conditions *without* necessarily following an SSW. If we examine the NAO anomalies from 5,000 random 30d periods that start in DJF, then the chance of them being negative is 43% (in both the model and the observations; this is below 50%, because the distribution of NAO anomalies is negatively skewed), with a 95% confidence interval in the model of 42%–45%. Our result of a 65% probability of a negative NAO response therefore represents a significant increase in the chance of a negative NAO month following an SSW.

The distribution of individual post-SSW NAO responses is shown in Figure 1(c), and the distribution of mean NAO responses (averaged over all SSWs) across the hindcast resamples is shown in Figure 1(d). There is a very broad range of possible NAO responses in the model, including positive as well as negative outcomes, although the distribution is strongly shifted towards negative NAO conditions. The observations and the model span very similar ranges, and have very similar mean responses (which are not significantly different). We can therefore say that the NAO response to SSWs in our model is indistinguishable from the observations.

Composite mean vertical profiles of the zonal mean zonal wind anomalies at 60° N are shown in Figure 2, following the similar plots of Baldwin and Dunkerton (2001). These illustrate the mean vertical progression of anomalies in the zonal mean circulation following an SSW, and again demonstrate the impact of the different sample sizes on the results. The observational composite is clearly "noisier" than the model data due to the limited number of events. However, the observations are broadly consistent with the model resamples: the only large areas where the observations lie outside the 95% range of the resamples (pink cross-hatching) are at lags > 60d, where there are very few observed events.

Both model and observations exhibit a near-surface easterly anomaly for at least the first 30 days following the SSW. This motivates our use of the 30d post-SSW period for assessing the NAO responses.

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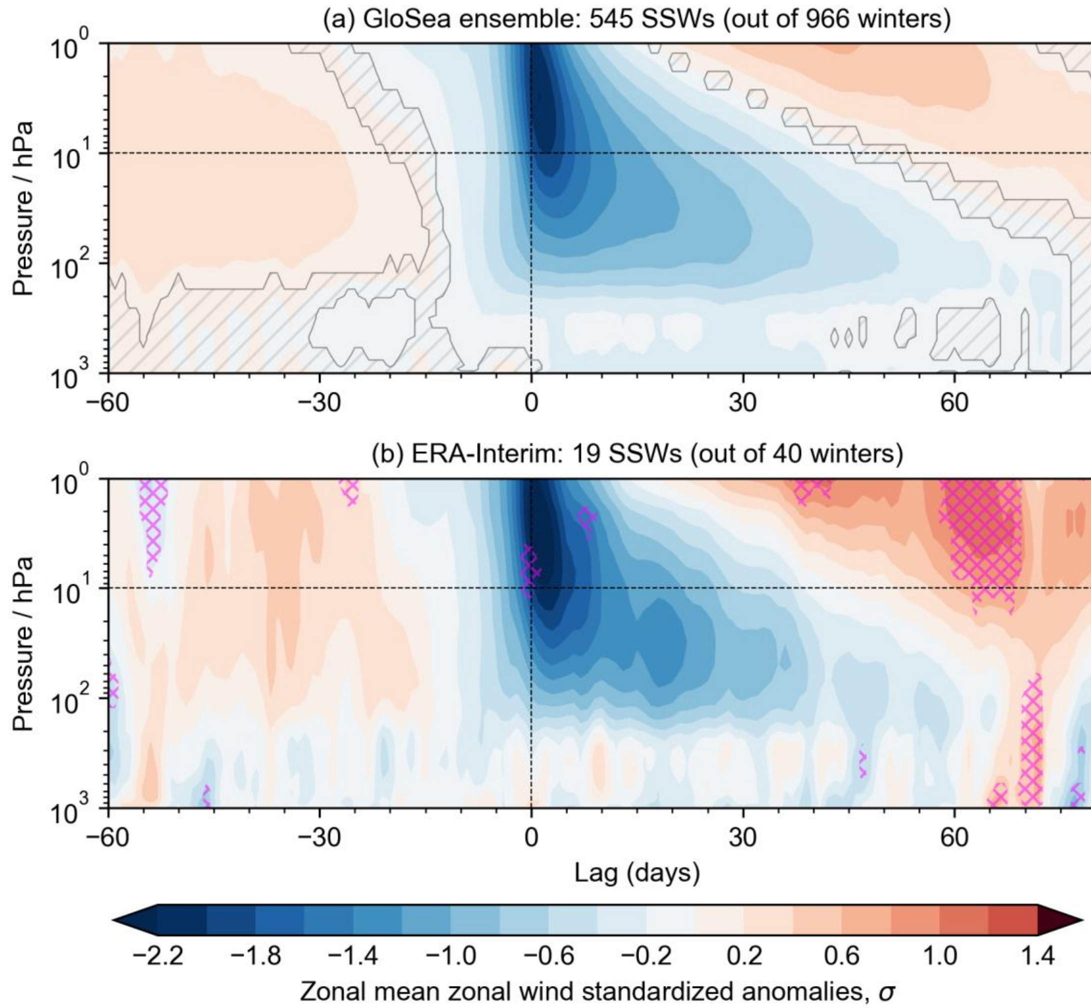

Figure 2. Development of zonal mean zonal wind anomalies associated with SSWs in (a) models and (b) observations. Both panels show composite mean standardized anomalies in the 60° N zonal mean zonal wind profiles, with each contributing winter centred at their SSW date (lag = 0). Only dates in DJFM contribute; the composite comprises fewer SSWs at large lags/leads. In the model panel (a), areas are hatched where the anomalies are indistinguishable from zero. In the observation panel (b), pink cross-hatching marks where the anomalies exceed the 95% range of the hindcast resamples.

Finally for this section, Figure 3 shows the PMSL composite means over the 30 days following SSWs. The mean negative NAO response shown in Figure 1(d) can be seen clearly here, and although the response is nominally weaker in the model, there is no statistically significant difference between model and observations in the North Atlantic or Arctic regions: the observations fall within the range expected from the model resamples.

A region in the observations covering central/eastern Europe, North Africa and the Middle East does however lie outside the model resamples' 95% range (purple cross-hatching), and a similar area shows a significant difference with the model mean (panel c). An enhanced Aleutian Low is seen in the model composite, which agrees with other model results (e.g. Ineson and

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Scaife, 2009), and with observations in response to SSWs in El Niño winters in particular (e.g. Butler et al., 2017). This region has the opposite sign in the observational composite (panel b), which agrees with others' observational results more generally (e.g. Butler et al., 2017; Oehrlein et al., 2021). This suggests that there may be some minor differences in the relative frequency, strength and/or persistence of El Niño/La Niña events or their PMSL precursors/responses in our model compared to our observational sample (recall that the observations cover a slightly longer period than the hindcast). However, although the peak of the high anomaly lies outside the model resamples' 95% range (purple hatching), the means are not significantly different to each other (panel c).

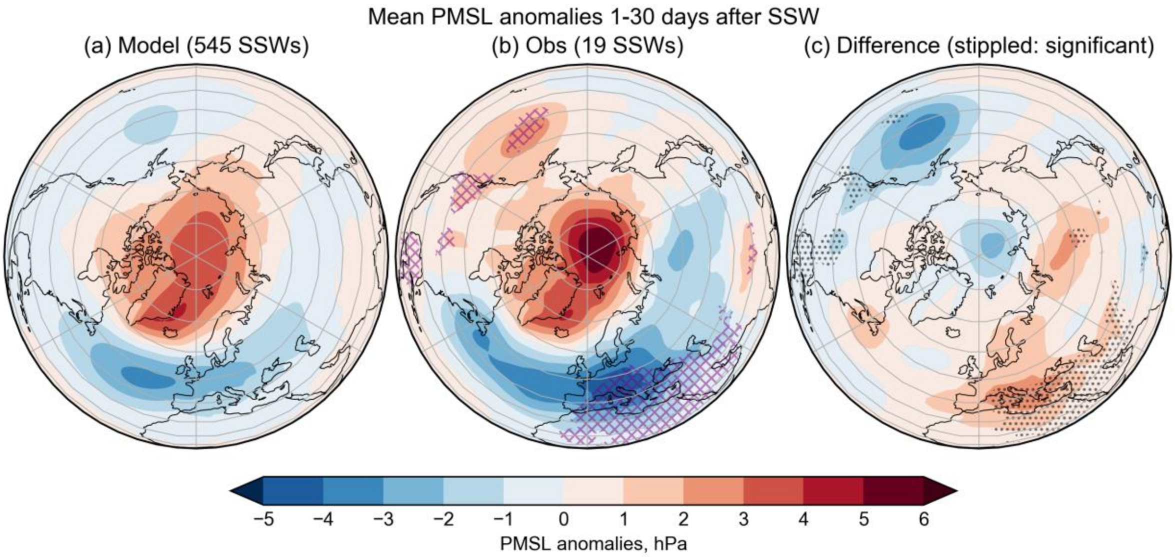

Figure 3. Surface responses to SSW events. The left and centre panels show the composite mean PMSL fields for (a) the model and (b) observations, in the 30d after an SSW. In the observational panel, regions where the data exceeds the 95% range of composites from the hindcast resamples are marked with purple cross hatching. The right-hand panel (c) shows the difference between the mean fields (model minus observations), with stippling indicating where the differences are statistically significant.

We can therefore conclude that the model is statistically indistinguishable from the observations for the features that we focus on in this study: the frequency of SSWs, the distribution and strength of post-SSW 30-day mean NAO responses, and the frequency of negative NAO responses.

4 What affects the NAO response to SSWs?

Having established the realism of our large model ensemble, we now assess atmospheric features that might have a robust impact on the NAO response.

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4.1 Tropospheric precursors of negative NAO responses

Following Figure 2, we use a 10d pre-SSW period to examine the impact of surface precursors on the 30-day post-SSW NAO response. We first note that the pre-SSW NAO itself is not a strong precursor of a negative NAO response. The correlation between the pre-SSW and post-SSW NAO anomalies is 0.19, which is statistically significant but very small. This is consistent with Christiansen (2005), who showed a similar result for the AO, and with Domeisen et al. (2020), who used a weather regimes approach to examine tropospheric evolution around SSWs. Another way of assessing the impact of a potential precursor is to use its terciles (for example) to select SSW events, and compare the resulting likelihoods of negative NAO responses in those subsets. If we pick SSWs according to the upper and lower terciles of the pre-SSW NAO, then the probability of a post-SSW negative NAO changes from 65% for the full sample (as in Figure 1b), to 66% and 72% respectively; these are not significantly different to each other.

We can identify other regions that might act as surface-level precursors of negative NAO responses, by mapping the correlation between pre-SSW PMSL and post-SSW NAO anomalies (Figure 4). The weak correlations seen in the model data emphasize the very high degree of variability from one SSW to another, and thus provide important context for how we interpret other results. Large areas of PMSL over the polar cap, the subtropical North Atlantic, and North Pacific show statistically significant correlations with the post-SSW NAO response. However, most of these regions are not statistically significant in the observational map. Two large areas in the extratropics that stand out as being statistically significant in the observations are in the Russian Far East north of the Sea of Okhotsk, and western Russia and Kazakhstan around the Ural Mountains. Although the model correlations show a similar area of significance for the far eastern region, the Ural region in the model contains large areas without significant correlation. If we consider area-averaged PMSL anomalies, the correlation of the Ural region pre-SSW with the post-SSW NAO in the model is very small (-0.14, although significantly different to zero). If we select SSW events according to the upper and lower terciles of the pre-SSW Ural PMSL anomalies, we find this changes the probability of negative NAO from 65% (as in Figure 1b) to 76% and 58% respectively, which are significantly different from each other.

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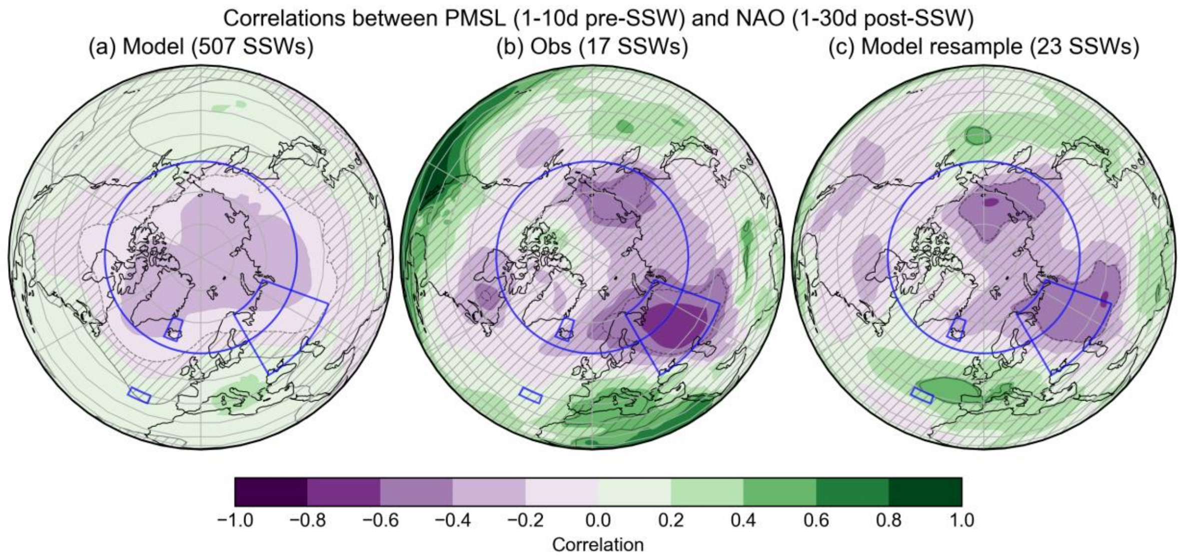

Figure 4. Correlation of surface precursors with the NAO response to SSWs. The left and centre panels show the correlations between the 10d pre-SSW PMSL fields and the 30d post-SSW NAO anomalies for (a) the model and (b) observations. The right-hand panel (c) shows the correlation map for a particular case from the hindcast resamples, chosen as the sample whose correlation map has the highest spatial correlation with the observed correlation map for points north of 40° N (0.73). In each panel, the 95% confidence intervals are marked with a contour, and areas where the correlation is indistinguishable from zero are hatched. Regions used in our analysis (the NAO boxes, a Ural region and the Polar cap) are outlined in blue.

If we instead pick a region covering the Polar Cap (PMSL anomalies averaged north of 60° N), we get a stronger relationship: the post-SSW NAO-negative probabilities are differentiated into 78% and 53% for the upper and lower terciles[1] of the pre-SSW PMSL anomalies respectively. The correlation is stronger too, at -0.34, which is not only significantly different to zero, but significantly stronger than that from the Ural region. High pressure or blocking, approximately over Eurasia (and often the polar cap), has often been found to be a precursor to SSWs in general (e.g. Cohen and Jones, 2011; Kolstad et al., 2010; Kolstad and Charlton-Perez, 2011). Our results extend this by quantifying the impact of these regions as precursors of post-SSW negative NAO conditions.

Our results demonstrate that just using the observations to select and test the impact of possible SSW precursors might result in a sub-optimal choice of areas. Using the model data allows more statistically robust signals to emerge from the noise, allowing us to be more confident in the results. Figure 4 also shows the correlation map from the 40-winter model resample that has the greatest *spatial* correlation with the observed correlation map north of 40° N, highlighting that the model can

[1] Although our choice of terciles is arbitrary, our results are qualitatively unchanged if we select by different quantiles. If we use the outer quintiles of the pre-SSW polar cap PMSL (resulting in smaller samples), the probabilities of negative NAO post-SSW are instead 82% and 49%.

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produce similar relationships to those seen in the observations; the differences with the full model map are therefore largely due to better sampling.

4.2 Zonal wind precursors

We extend the above approach to examine how the post-SSW NAO response changes when we subselect SSWs according to the 60° N zonal mean zonal wind (ZMZW) anomalies at different heights, and different lead times, prior to the SSW. The impact on the probability of a negative NAO response is shown in Figure 5, in terms of the difference between the probabilities from picking lower and upper terciles of the ZMZW on each pressure-level; a positive difference indicates an NAO-negative response is more likely for the SSWs preceded by more easterly ZMZW anomalies at a given height and lead-time. For the immediate 10-day pre-SSW period (rightmost column in Figure 5), as used in the previous section, the ZMZW in the lower stratosphere and troposphere can clearly affect the probability of a negative NAO response: if we pick SSWs in the upper and lower terciles of the ZMZW at 100 hPa, then the subsequent NAO-negative probabilities are 56% and 73% respectively. The correlation of the ZMZW at 100 hPa with the post-SSW NAO is 0.30 (not shown), which is also statistically significant.

In contrast, if we focus on the 10 hPa level (which we use to define whether or not there is an SSW), there is no significant relationship between the 10d pre-SSW values and the post-SSW NAO: selecting by terciles only changes the NAO-negative probabilities to 61% and 67% (not significantly different), and the correlation with the NAO is 0.09 (also not significant).

Figure 5 also shows how these features vary as we examine earlier pre-SSW periods. Although there are some signs that the 10 hPa ZMZW can be a significant precursor around 2–3 weeks before the SSW, the stronger signals remain robustly in the lower stratosphere: e.g. terciles of the 21–30d pre-SSW ZMZW at 100 hPa can still separate the post-SSW negative NAO probabilities into 61% and 75%. The relationship between the pre-SSW lower stratosphere and the post-SSW NAO remains significant for 10-day periods as early as 3–4 weeks before the SSW. Therefore selecting SSWs according to the zonal mean zonal winds in the lower stratosphere is a robust way of selecting events that are more/less likely to result in a negative NAO response.

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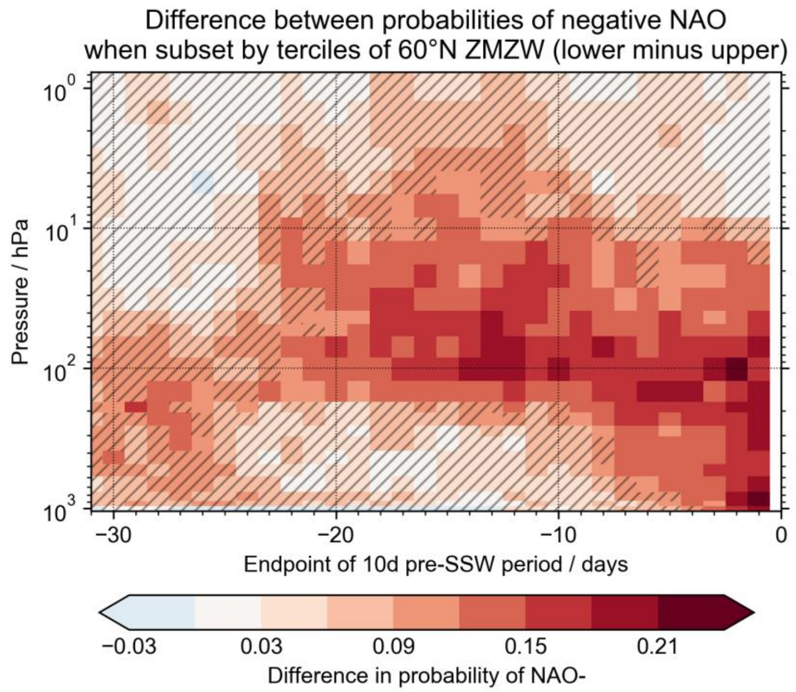

Figure 5. Effect of zonal mean zonal wind (ZMZW) precursors on the NAO response to SSWs. The shading shows the difference in the probability of a negative NAO response, between selecting lower and upper terciles of the ZMZW, at each pressure level and pre-SSW lead time. Points are coloured on the final day of the 10d pre-SSW period they use. Points are hatched out when the difference is statistically indistinguishable from zero.

4.3 Zonal wave precursors

Here we separate our sample of SSW events according to whether they are wave-2 dominated or wave-1 dominated, based on the ratio of amplitudes of the first two zonal wavenumbers, A_2/A_1, in the 60° N eddy geopotential height at 50 hPa on the day of the SSW. An SSW is considered wave-2 dominated if $A_2/A_1 > 1$, and wave 1 dominated if $A_2/A_1 < 1$. The frequency of SSWs that are wave-2 dominated, and the difference in NAO responses between wave-2 and wave-1 dominated SSWs, are shown in Figure 6. The model has 140 out of 545 SSWs that are wave-2 dominated (26%, with a 95% confidence interval of 22%–30%). The proportions in the observed data are very similar, which means only 6 out of 19 SSWs (32%) are wave-2 dominated. The central 95% range of the model resamples is 9% to 45%, which we take to be a measure of the uncertainty in the observed frequency. This again illustrates that the model is statistically indistinguishable from the observations in this regard, and emphasizes that the observations are highly uncertain.

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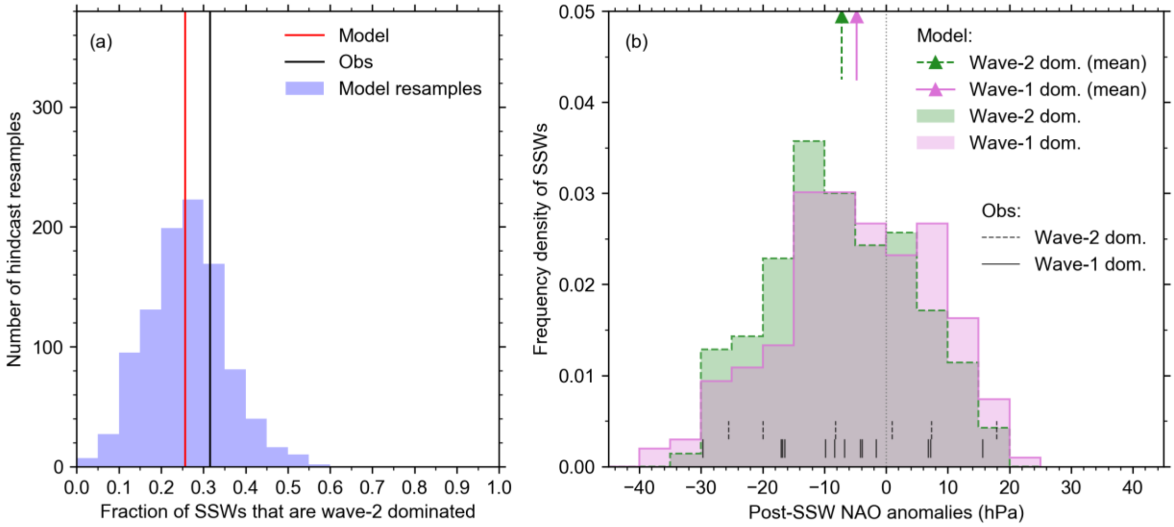

Figure 6. Frequency of wave-2 dominated SSWs, and their NAO responses. Panel (a) shows the frequency of wave-2 dominated SSWs in the model and observations (vertical lines), and the distribution in the model resamples (histogram), in a similar format to Figure 1(b). Panel (b) shows the normalised distribution of NAO responses across the wave-2 and wave-1 dominated SSWs (as labelled), in a similar format to Figure 1(c). The distributions in the model are shown as histograms, and observations are shown as vertical tick marks near the base of the plot, vertically separated for clarity. The mean responses in the model are marked with coloured arrows at the top of the plot.

We have also tested using the wave amplitudes in the eddy geopotential height at 10 hPa instead of 50 hPa. However, as shorter wavenumbers are less able to penetrate to greater heights, following the Charney–Drazin theorem (Charney and Drazin, 1961), there are systematically fewer wave-2 dominated events by this definition: 82 out of 545 events in the model (15%), and correspondingly 3 out of 19 in the observations (16%). These fractions in the model and observations are again statistically indistinguishable, but rather than limit ourselves by these smaller numbers, we prefer to use the 50 hPa level to give a better representation of wave-2 prominence in the stratosphere.

Figure 6(b) shows that the NAO responses to both wave-2 and wave-1 dominated SSWs cover a similar range. The wave-2 dominated SSWs have proportionally more negative NAO responses, and fewer positive NAO responses, than the wave-1 dominated SSWs. However, the probabilities of a negative NAO response in the wave-1 and wave-2 dominated cases (63% and 72% respectively) are not significantly different.

The correlation of the wave amplitude ratio (in terms of $\log_{10} A_2/A_1$, as the ratio itself is highly skewed) with the post-SSW NAO is -0.07, which is also not significant. However, the mean NAO responses in the model are significantly different. These are shown in Figure 6(b), and can also be seen in a wider context in Figure 7, which shows maps of the composite mean PMSL response to wave-1 and wave-2 dominated SSWs: There is a clear negative NAO response in both cases, but the response in the wave-2 dominated case is stronger. Overall, although we can discern an impact of selecting wave-2 vs wave-1 dominated

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SSWs on the NAO response *on average*, this is much weaker than the other precursors that we have investigated: the wavenumber characteristics of an SSW are not a strong predictor of its NAO response.

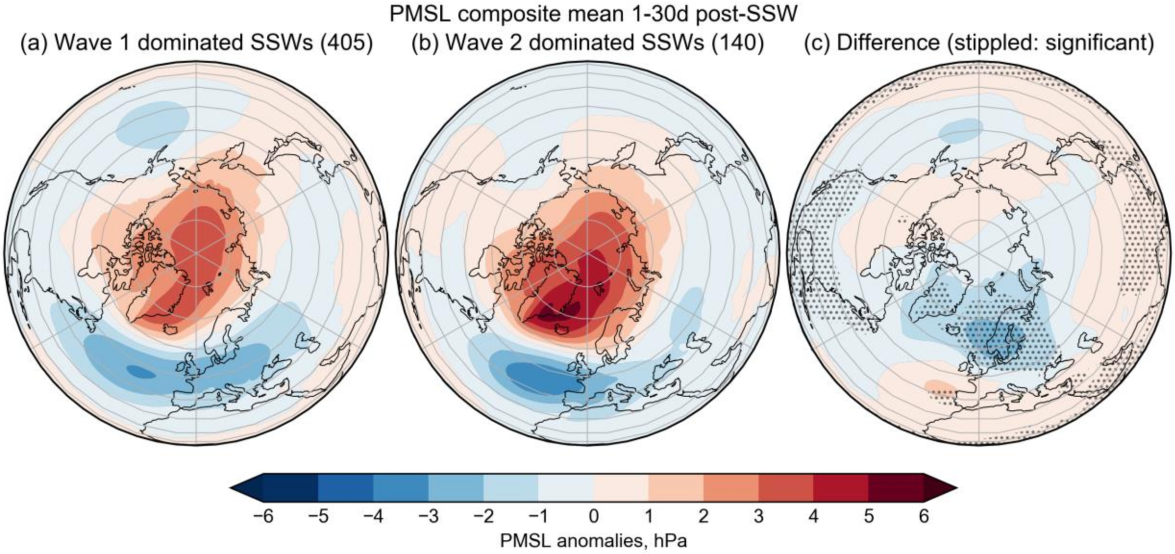

Figure 7. Surface response to wave-1 and wave-2 dominated SSWs. The left (a) and centre (b) panels show the composite mean PMSL anomalies 30d post-SSW, for the two sets of SSW events. The right-hand panel (c) shows the difference (wave-1 dominated cases minus wave-2 dominated cases), with stippling indicating where the differences are statistically significant.

5 Discussion and summary

We have shown that our ensemble of initialised climate model simulations is statistically indistinguishable from the observations in terms of the frequency of SSWs, the distribution of their NAO responses, and the proportions that are wave-1/wave-2 dominated. This has enabled us to determine features of the climate prior to an SSW that have a statistically significant effect on the probability of subsequent negative NAO conditions. The size of our ensemble, and its agreement with the observations, has meant that we have been able to do this more reliably and robustly than if we had based our assessments on the observations alone.

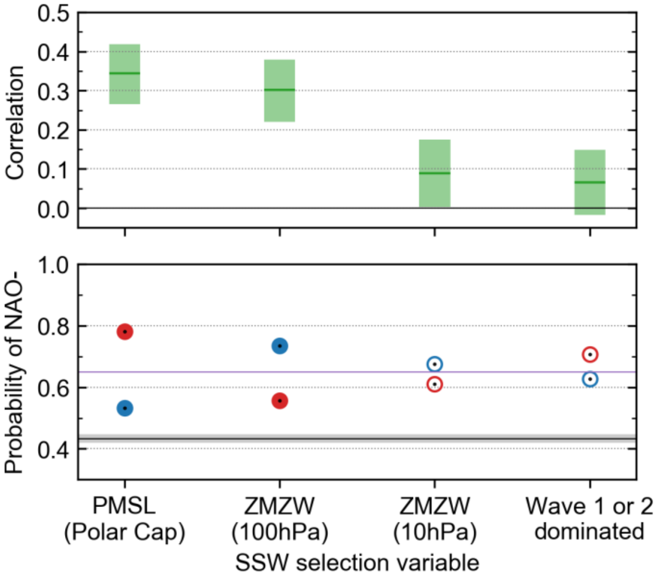

Figure 8. Summary of the impact of different factors on the NAO response to SSWs. Four key results are highlighted: the best-performing 10d pre-SSW PMSL region, the 10d pre-SSW 60° N zonal mean zonal wind at two different heights, and whether the SSW is wave-1 or wave-2 dominated at 50 hPa (in terms of $\log_{10} A_2/A_1$). Top: Correlation between each factor and the post-SSW NAO anomaly, with 95% confidence intervals. Correlations for the PMSL and wavenumber factors have been inverted for ease of comparison. Bottom: Changes in the probability of a negative NAO, after selecting upper (red) or lower (blue) terciles of each precursor, or if the SSW is wave-2 (red) or wave-1 (blue) dominated. Filled circles indicate that the two probabilities are significantly different to each other. The overall rate of negative NAO responses to SSWs (as in Figure 1d) is marked with a horizontal purple line. The black line with 95% confidence interval shading gives the climatological frequency of negative NAO conditions in random 30d periods that start in DJF.

Figure 8 shows a summary of our key results, in terms of correlations with the NAO response, and shifts in the probability of negative NAO responses, under different conditions. We have been able to rule out some features as being significant or strong determiners of the NAO response to an SSW: whether the SSW is wave-2 or wave-1 dominated, the magnitude of the prior NAO state, and the strength of the 60° N zonal mean zonal wind at 10 hPa before the SSW, all appear to have little bearing on the subsequent NAO state.

In contrast, the PMSL anomalies over the polar cap, and the ZMZW around 100 hPa, both averaged 1-10d pre-SSW, have a significant impact on the likelihood of subsequent negative NAO conditions: they can change the probability from the baseline value of about 2/3, down to around 55% or up to around 75%, when selecting by terciles of the precursor. However, even the reduced probabilities are greater than the climatological probability of 30-day negative NAO conditions (43%), so the presence of an SSW always increases the NAO-negative probability even given these modulating factors. The strongest correlations we have identified with the post-SSW NAO are around 0.3. With our large sample size, these are statistically significant, but it is important to emphasize that they are still relatively weak correlations.

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The observed correlations (not shown) are all consistent with the model values, as the small sample size ensures very wide confidence intervals. However, the significance tests also agree with the model results. For the Polar Cap PMSL precursor for example, the observed correlation between that and post-SSW NAO is -0.61, with confidence intervals of -0.84 to -0.18, covering the model value of -0.34 (although as discussed earlier, the eye is more likely to be drawn to the high apparent correlations in the Ural region than the polar cap as a whole, for the observations). For the correlation of the wave amplitude ratio with the post-SSW NAO, the observed value (0.19) is nominally the opposite sign to the model result (-0.07), but again the wide confidence intervals of -0.29 to +0.59 easily cover zero and the model value; and the model correlation is also not statistically significant.

The precursors we have identified are in broad agreement with other studies. Xu et al. (2022) found that SSWs preceded by a weaker polar vortex in the middle stratosphere, or with a more intense breakdown of the vortex, resulted in more robust surface impacts (a negative NAO pattern). They also found that the relationship with wave driving was not simple: the weaker prior vortex was mostly associated with wave-1 anomalies, whereas a more intense SSW was related to both wave-1 and wave-2 driving. This suggests that the impact of different wavenumbers on the evolution of the SSW could depend on the proportions of these different kinds of events in the available samples, and could explain the somewhat mixed results for wavenumber amplitude impact on the NAO that we have seen.

Karpechko et al. (2017) and Oehrlein et al. (2021) showed that SSWs with stronger anomalies in the lower stratosphere were more likely to have a surface impact, and that the state of the polar vortex at 10 hPa has very little relationship with the NAO response, in agreement with our results.

White et al. (2019) identify a region in the 700 hPa GPH as a precursor of downward-propagating SSWs, in a similar location to the Ural area we identified in observations as a potential precursor of a negative NAO response (Figure 4b). Although they also use a large model ensemble, their methods are considerably different to ours, and it is not clear how comparable their results are. They also find that the strength of upward wave flux is a related but stronger predictor of downward propagation of SSWs than tropospheric features like the Ural High. This could be an important area for future research.

The impact of unpredictable internal variability in masking the potential impact of precursors on the surface response is also frequently noted in the literature (e.g. Hitchcock and Simpson, 2014; Oehrlein et al., 2021; White et al., 2019), and is reflected in the low correlations we have seen.

While we have focused on precursors of surface impacts in the pre-SSW PMSL and ZMZW fields, there are many other features of the climate that affect the behaviour and evolution of SSWs. It has been long known that the stratospheric polar vortex is weaker, and SSWs more frequent, when the quasi-biennial oscillation (QBO) in the tropical stratosphere is in its easterly phase (e.g. Anstey and Shepherd, 2014 and references therein). The El Niño–Southern Oscillation (ENSO) affects the stratosphere, and El Niño events have been linked to an increased likelihood of SSWs (e.g. Domeisen et al., 2019 and references therein). And phases 6 and 7 of the Madden–Julian Oscillation (MJO), i.e. enhanced convection in the tropical west Pacific, have also been linked to SSWs (Schwartz and Garfinkel, 2017), both in terms of helping to trigger the event in the first place, and in making surface impacts such as a negative NAO or AO pattern more likely. Although the effects of these different driving phenomena are difficult to disentangle, there have been some promising results (e.g. Liu et al., 2014; Ma et al., 2020). As with other SSW-related studies, the small observational sample is a limiting factor when drawing robust conclusions; but it also suggests that this area would be an interesting target for further research using the hindcast-based approach we have demonstrated here.

We hope that our findings, based on features of the surface pressure and polar vortex winds, will help to clarify forecast assessments based on the state of the climate system prior to SSW events. Although precursors exist, we emphasize that they only have a modest influence on the probability of an SSW being followed by a negative NAO and its attendant impacts.

Data availability

ERA-Interim data is available from ECMWF at https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era-interim. Although we used GloSea5 hindcast data available internally at the Met Office, much of the data we used is also available from the Copernicus Climate Change Service Climate Data Store, at https://doi.org/10.24381/cds.181d637e (surface data) and https://doi.org/10.24381/cds.50ed0a73 (data on pressure levels); system codes 12 and 13 correspond to the hindcast runs used here.

Author contribution

PB conducted the analysis and wrote the original draft. AS, SH and HT supervised and contributed to the interpretation of the results. All authors contributed to the writing, reviewing and editing of the manuscript.

Competing interests

The authors declare that they have no conflict of interest.

Acknowledgements

PB would like the thank Lesley Gray and Mark Baldwin for helpful discussions during the early stages of this study. This work and its contributors (PB, AS, SH) were supported by the UK–China Research & Innovation Partnership Fund through the Met Office Climate Science for Service Partnership (CSSP) China as part of the Newton Fund. SH and HT were supported by the Met Office Hadley Centre Climate Programme funded by BEIS.

https://doi.org/10.5194/wcd-2022-39
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