Predictability of European Winters 2017/2018 and 2018/2019: Contrasting influences from the Tropics and stratosphere

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Abstract
The European winters of 2017–18 and 2018–19 were not climatically extreme, but both winters had a major sudden stratospheric warming (SSW). In February 2018, an SSW led to an intense cold outbreak across Europe and further spells of cold weather in March. The SSW of January 2019, although well predicted and expected to increase the chance of a cold end to winter, apparently produced little impact. In this study, we examine the performance of the Met Office seasonal prediction system in these winters, and the influences that led to these outcomes. To achieve this latter objective, sets of numerical experiments are performed in which the tropical troposphere and the extratropical stratosphere are relaxed towards their observed state, allowing the influence of each on the North Atlantic-European atmospheric circulation to be identified. Using these experiments, we show that the SSWs had similar impacts in each case, creating a signal of easterly surface wind anomalies in the weeks following the event. In contrast, tropical influences were opposite in the two winters, acting to strengthen the easterly signal at the end of February 2018 and opposing it in January 2019. The different apparent responses to the two events therefore came about largely through tropical tropospheric variability. Furthermore, we highlight the importance of a very strong cycle of the Madden-Julian Oscillation (MJO) in late January and early February 2018 as an important driver for the February 2018 SSW. MJO teleconnections appear to have been critical in creating the large mid-latitude wave 2 amplitude that has been identified as the immediate cause of this event.

KEYWORDS
European climate, seasonal prediction, sub-seasonal prediction, sudden stratospheric warmings, teleconnections, winter 2017–18, winter 2018–19, winter North Atlantic oscillation
1 | INTRODUCTION

Skilful numerical seasonal prediction of European winters (Scaife et al., 2014; Athanasiadis et al., 2017) is possible through prediction systems’ ability to represent the effect of systematic external influences on the regional atmospheric circulation. These influences can be persistent, resulting in changes in the occurrence of specific types of weather lasting well beyond the approximately 10-day timescale of ‘weather noise’ (Feldstein, 2000). Their sources reside in other parts of the climate system, which are linked to the region via atmospheric teleconnections (Liu and Alexander, 2007; Nigam and Baxter, 2015; Stan et al., 2017). For example, the multi-seasonal timescale of large El Niño or La Niña events can provide an influence throughout a European winter (Moron and Gourrand, 2003; Brönnimann, 2007; Ineson and Scaife, 2009; Li and Lau, 2012a, 2012b; Jiménez-Esteve and Domeisen, 2018).

Many other slowly-varying sources of predictability for European winters have been identified, including North Atlantic sea-surface temperatures (SST; Czaja and Frankignoul, 2002; Rodwell and Folland, 2002), conditions in the wider tropics (Scaife et al., 2017a), and the 11-year solar cycle (Kodera, 2002; Ineson et al., 2011; Chiodo et al., 2012; Andrews et al., 2015; Gray et al., 2016).

A particularly important influence on European winters is from the stratosphere. In winter, the Arctic stratospheric polar vortex (SPV) is dynamically coupled to the tropospheric circulation (Perlwitz and Graf, 1995), especially over the North Atlantic sector (Scaife et al., 2005). Predictability from the SPV arises from both its present state (Nie et al., 2019) and the fact its strength is subject to influences from long-lived sources such as the El Niño-Southern Oscillation (ENSO; Domeisen et al., 2019) and the Quasi-Biennial Oscillation (QBO; Holton and Tan, 1980; Andrews et al., 2019). Both strong and weak SPV states can influence tropospheric circulation, with strong events tending to enhance the tropospheric zonal flow and weak events tending to reduce it (Baldwin and Dunkerton, 2001). Major sudden stratospheric warming (SSW) events, in which the SPV is entirely displaced or disrupted (Butler et al., 2015), are frequently followed by impactful spells of cold weather in Europe associated with easterly winds (Scaife and Knight, 2008). Nevertheless, not all SSWs result in such spells, and it remains unclear whether systematic differences between SSWs (Mitchell et al., 2013; Karpechko et al., 2017) or unpredictable tropospheric variability (Afargan-Gerstman and Domeisen, 2020) account for apparent differences in surface response.

The European winters of 2017–18 and 2018–19 were generally unexceptional overall (on average over the December–January–February period of each year). Winter 2017–18 was generally colder than the winter of 2018–19 but was still close to the 1981–2010 long-term average temperature in many parts of the continent. Despite this, both winters saw a major SSW event, occurring in mid-February in 2018 and in early January in 2019. Following the 2018 SSW, easterly winds were established over much of the continent, leading to an intense and widespread cold episode in late February and early March. Further intermittent cold was experienced throughout most of Europe through March 2018. In contrast, only brief and less widespread cold occurred in Europe in early 2019, despite the occurrence of an SSW that was at least as strong as that of the previous year.

In this study we examine how well these two winters were predicted in the Met Office seasonal prediction system, following on from studies of system performance in earlier winters (Scaife et al., 2017b; Dunstone et al., 2018). We also analyse reanalyses of observations and model simulations to infer the external influences that were acting on the North-Atlantic European circulation in each winter. Our focus is on understanding the apparent difference in impacts of the two SSWs that occurred these winters.

2 | METHODS

ERA-Interim reanalysis (Dee et al., 2011) is used in place of observational data for the two winters. Synoptic and monthly mean data were obtained for a range of key diagnostic fields. Dates of SSW events are defined as the calendar day for which the zonal mean zonal wind at 10 hPa and 60°N is westerly at 0UT but easterly at 0UT of the following day. Synoptic data are also averaged to provide daily and calendar week (Monday to Sunday) means. We define an index of the North Atlantic Oscillation (NAO) as the difference in the pressure at mean sea level (PMSL) between the Azores (25.70°W, 37.74°N) and south west Iceland (22.61°W, 63.98°N). Data for the principal components of the Madden-Julian Oscillation (MJO) (Wheeler and Hendon, 2004) were exceptionally obtained from the Australian Bureau of Meteorology (http://www.bom.gov.au/climate/mjo/graphics/rmm.74toRealtime.txt).

Real-time seasonal forecasts were produced with the Met Office GloSea5 long-range prediction system (MacLachlan et al., 2015). Forecasts were initialized with atmospheric and oceanic analysis data between the 30th (29th) October and 19th (18th) November 2017 (2018), with two ensemble members each day making a lagged
ensemble of 42 members for each winter. Each member was integrated for approximately 7 months to simulate the winter–spring periods. In addition, a set of hindcasts for the winters 1993–4 to 2015–16 was produced. Seven members were initialized on each of the 1st, 9th, 17th, and 25th of each month giving a total of 28 members per winter. They are combined using the method of MacLachlan et al. (2015) to correct each day’s forecasts for time-dependent biases. Pattern correlations \( r \) between GloSea5 and ERA-Interim North Atlantic-European PMSL are calculated using area-weighting over the domain 75°W–35°E, 25°–70°N.

Further experiments are performed post-event to analyse the global influences on the European region in the two winters. We produce GloSea5 ensembles in which part of the atmospheric model is constrained to be similar to the atmospheric state in the ERA-Interim reanalysis, while the rest of the model evolves according to its own dynamical and physical tendencies. Specifically, the atmospheric variables \( u \), \( v \), and \( T \) are relaxed towards their reanalysis values within the specified domain with an exponential timescale of 6 hr (Knight et al., 2017; Maidens et al., 2019). Two sets of experiments are performed for each winter–spring period: one in which the tropical troposphere (19.5°N to 19.5°S; surface-16 km altitude) is relaxed, and another in which the extratropical stratosphere (41.5°-90°N; 20–85 km altitude) is relaxed. Additional linear tapering of the relaxation strength (the inverse of the timescale) is applied over 5.5° latitude at the edges of the domain to reduce any shock from the relaxation. Initial conditions for both winters are taken from 1st November of the years 1993–2015 (three members per year, making a total ensemble size of 69) rather than the year in question. This allows us to remove (to first order) the effect of initialisation on the ensemble mean of the relaxation experiments. Additional ensembles are performed using relaxation data for each of the 23 hindcast years (three members per year) to provide a reference for the calculation of anomalies. Subtracting the means of these ensembles corrects for any systematic effect of the relaxation and isolates the response of the extratropical troposphere to anomalies in the imposed relaxation data.

Note that the relaxation data used in the stratospheric relaxation experiment may contain tropically forced signals. The experiments are not, therefore, intended to partition tropical and non-tropical influences. As in previous studies (e.g., Knight et al., 2017), the use of large ensembles (both from GloSea5 and the relaxation experiments) ensures that the simulated features are significant to a high level of confidence.

### 3 | RESULTS

#### 3.1 | Winter forecast

Forecast ensemble mean PMSL averaged over the winters of 2017–18 and 2018–19 are compared to reanalysis in Figure 1. Although ensemble predictions are intended to give probabilities of potential scenarios, the ensemble mean is a useful summary of forecast signals. For winter 2017–18, the reanalysis shows a zonal PMSL dipole in the mid-latitude North Atlantic which resembles the positive phase of the NAO. The winter mean NAO index anomaly is +6.7 hPa. GloSea5 predicted the North Atlantic pattern well (\( r = 0.60 \)) with an ensemble mean NAO index value of +3.7 hPa. This value is clearly smaller than seen in the reanalysis but is significant compared to the interannual SD of the hindcast ensemble mean (2.5 hPa). Elsewhere, the strong high-pressure anomaly in the Aleutian region is also well predicted. Both Atlantic and Pacific features are consistent with the weak La Niña event that was underway during winter 2017–18, showing notable similarity to GloSea5’s composite La Niña response for January–February (Ayarzagüena et al., 2018). La Niña has been shown to act to fill the Aleutian cyclone (Taguchi and Hartmann, 2006) and favour positive winter mean NAO (Gourand and Moron, 2003; Iza et al., 2016).

Winter 2018–19’s North Atlantic PMSL anomalies show no dominant large-scale pattern. This reflects different PMSL patterns in its constituent months (see Section 3.4), with a winter NAO close to neutral (−2.2 hPa). GloSea5’s predicted patterns over the North Atlantic region show little resemblance to reanalysis (\( r = 0.16 \)). Its ensemble mean NAO index anomaly was negative (−1.0 hPa), albeit by less than 1 SD (2.5 hPa). The predicted patterns are likely linked with the weak El Niño underway in November 2018, although are not as clearly a result of tropical Pacific conditions as the previous winter. Differences with the observed patterns may be linked to the lack of negative NAO following that winter’s SSW, as shown below.

#### 3.2 | February 2018 SSW and European cold spells

A major SSW occurred on 11th February 2018. This was a wave 2 driven vortex splitting event (Lü et al., 2020). GloSea5 was among the leading sub-seasonal to seasonal prediction systems to predict the SSW (Rao et al., 2020). An increasing probability of a weak SPV was seen in
forecasts initialized during the last week of January 2018, with deterministic predictability of the SSW’s occurrence appearing at a lead time of 11 days (Karpechko et al., 2018). In December, the SPV had been weaker than average but became stronger than average in January, leading up to the SSW (Figure 2). After the SSW, easterly anomalies propagated downwards reaching the surface in late February, which corresponds to the initial cold spell over Europe. The SPV remained weak into April and while anomalies in the tropospheric zonal flow were intermittent, further anomalously easterly episodes in mid-March and at the start of April each resulted in cold European spells.

The GloSea5 forecast for December–February 2017–18 indicated a slight signal for a weak SPV at 10 hPa (Figure 2), but unusually this appears disconnected from the SPV in the lower stratosphere, which was predicted to be stronger than usual, consistent with the positive NAO signal. The tropical tropospheric relaxation ensemble shows that the effect of the Tropics was generally to strengthen the SPV. The mid-winter period, with a strong SPV, is reproduced. The SSW is not simulated in the ensemble mean, and taken at face value, these results suggest that the SSW was unlikely to be forced by the Tropics. Nevertheless, we will show in further analysis that this may not be the case (see below). Note also that despite the absence of the SSW, an easterly anomaly appears in the troposphere in the tropical relaxation ensemble mean in late February at the time of the European cold spell. This implies that tropospheric teleconnections from the Tropics contributed to this event.

The tropospheric response to the stratospheric winds imposed in the extratropical stratospheric relaxation simulations shows an easterly anomaly at the surface in late February. This confirms the results of relaxation experiments by Kautz et al., 2020 showing the SSW contributed to the European cold spell. Furthermore, the near-surface flow until April and even into May is biased easterly, showing the SSW influence lasted well into spring. Surface easterly anomalies are not seen continuously during the post-SSW period in observations, because of unpredictable weather variability. The simulations show that the forcing of the troposphere by the 2018 SSW was similar to canonical estimates of SSW effects averaged over many events (Baldwin and Dunkerton, 2001).

The surface weather patterns leading up to and following the SSW are shown in Figure 3. For the week of

**FIGURE 1** PMSL anomalies (hPa) for the winters (December to February) of 2017–18 (top row) and 2018–19 (bottom row). ERA-Interim reanalysis anomalies (with respect to a 1981–2010 climatology) are shown in the left-hand panels. Real-time predictions made with GloSea5 prior to the season are shown in the right-hand panels. GloSea5 data are the difference between forecast and hindcast ensemble means. Note the nonlinear colour scale.
the 5–11 February (immediately prior to the SSW), the PMSL pattern resembled the positive phase of the NAO in the North Atlantic sector, with high latitude blocking further downstream over northern Eurasia. This blocking has been identified as a pre-cursor to the SSW (Karpechko et al., 2018), contributing to a large tropospheric wave 2 amplitude and acting as a source of upward-propagating planetary waves. Anticyclonic anomalies also occurred in the eastern North Pacific, further leading to a large wave 2 amplitude. The relaxation experiments indicate that the North Atlantic-Eurasian pattern appears to result primarily from stratospheric influences ($r = 0.78$), consistent with a stronger-than-usual SPV increasing the likelihood of positive NAO (Scaife et al., 2015). Despite this, neither the stratospheric nor tropical influences alone appear to explain the Eurasian block, which has been linked to snow cover anomalies (Lü et al., 2020) or could be intrinsic mid-latitude variability. In contrast, the Aleutian anticyclonic anomaly is clearly associated with tropical forcing, which explains almost all the features of the PMSL pattern in the eastern Pacific. Note that the tropical relaxation experiment does not reproduce the full amplitude of the North Pacific feature. As in the GloSea5 forecast, ensemble mean extratropical variability is too weak as a result of the signal-to-noise problem (Eade et al., 2014; Baker et al., 2018; Scaife and Smith, 2018).

In the first week following the SSW (12–18 February), a negative Arctic Oscillation pattern appeared, consistent with the response to the disruption of the SPV seen in the stratospheric relaxation experiment (and inconsistent with the response to tropical forcing seen in the tropical relaxation experiment). In the second week (19–25 February), high pressure intensified to the north of Eurasia, again consistent with the stratospheric relaxation experiment. Tropical relaxation, however, now also shows a negative NAO pattern ($r = 0.47$), which (in common with the stratospheric influence—$r = 0.53$) can explain the low-pressure anomalies over the Mediterranean extending into the North Atlantic. The experimental design, and the fact that the tropical experiment does not produce an SSW-like weakening of the SPV, means that this signal follows a tropospheric pathway from the Tropics. It corresponds to the initial stages of the signal seen in the tropical relaxation zonal mean wind in Figure 2. The culmination of the cold spell occurred in week 3 (26 February—4 March), by which time a strong negative NAO had emerged. Both the
tropospheric \( r = 0.77 \) and stratospheric \( r = 0.92 \) signals align closely with this pattern. In subsequent weeks (not shown), agreement between observations and the tropical relaxation experiment disappears, while stratospheric relaxation continues to mirror the observed negative Arctic Oscillation (AO)/NAO pattern that brought further cold spells later in March. During the late February cold spell, therefore, shorter-lived tropical forcing added to the longer-lasting forcing from the stratosphere to make the easterly outbreak particularly strong.

### 3.3 Tropical forcing of the February 2018 SSW

The strong tropospheric wave 2 pattern in the top left panel of Figure 3 was responsible for the SSW in February 2018 (Lü et al., 2020). We showed above that this pattern originated in part from tropical influences. Pacific PMSL patterns leading up to the event are indicative of Rossby wave propagation generated by tropical sources. This process is familiar as the means by which the winter climatological Aleutian cyclone deepens and fills in response to El Niño and La Niña, respectively (Yeh et al., 2018). Despite consistent La Niña conditions in winter 2017–18 (each winter month had a Niño 3.4 amplitude of just below 1 K), the positive North Pacific PMSL anomaly was strongest in February. Additional forcing came from a very strong Madden-Julian Oscillation (MJO) cycle that developed in January 2018 and propagated into the Pacific towards the end of that month. The MJO has been shown to increase the likelihood of both the development of SSWs (Garfinkel et al., 2012) and the negative phase of the NAO.
(Cassou, 2008) after it has passed into the tropical Pacific. In particular, phases 6 and 7 of the MJO (Wheeler and Hendon, 2004) are linked with tropospheric ridging in the northeast Pacific as a result of production of Rossby waves in response to MJO’s large-scale convective anomalies. The SSW event of January 2009 has been linked to the MJO via similar extratropical circulation patterns (Schneidereit et al., 2017), although MJO activity in 2009 was not as extreme as in early 2018. As in most prediction systems, the MJO in GloSea5 is not predictable beyond the sub-seasonal timescale (Lim et al., 2018) and was not predicted by the GloSea5 forecasts discussed here.

In late January 2018, the MJO’s projection onto phases 6 and 7 grew very strong (Figure 4, upper panel). It remained strong (if not at peak strength) until mid-February. It is during the time MJO phase 6/7 was strong that positive North Pacific PMSL anomalies appeared. In agreement with this, the amplitude of the tropospheric wave 2 started to increase at the beginning of February (Figure 4, second panel). This is several days after the growth of MJO phase 6/7, as expected for Rossby wave propagation to the extratropics (Jin and Hoskins, 1995). A peak amplitude of over 250 m was reached in the second week of February. A positive node of wave 2 was very well aligned with the location of Pacific anticyclonic anomaly throughout the time it had a high amplitude (Figure 4, second panel). The blocking over Eurasia was generally much less well aligned with wave 2, however, except for a period of about a week from 7th February (Figure 4, second panel). This suggests that the wave 2 forcing was initially a result of the Pacific anticyclonic anomaly, but that a blocking event subsequently developed over Eurasia, acting to intensify it.

The polar vortex started to weaken at the end of the first week of February (Figure 4, third panel). This is before the Eurasian anomalies were contributing to wave 2 (note wave 1 was not particularly strong at this time), suggesting that the initial vortex weakening is a result of the Pacific anomaly. Following consistently elevated wave 2 forcing, the vortex undergoes a split-type SSW on 11th February. Zonal mean winds remained easterly until the start of March, allowing a lengthy window of influence on the troposphere. Following downward propagation of the SSW signal, the NAO index started to decline soon after mid-month and became strongly negative (~ −50 hPa) by the last few days of February (Figure 4, lower panel), implying intense easterly flow over Europe. Note that despite imposing the large MJO event in the tropical relaxation ensemble, the model does not

**FIGURE 4** Time series of key variables from reanalysis leading up to the European cold spell in late February 2018. (a) Projection of the MJO vector in the space of its rotated principal components onto the unit vector bisecting MJO phases 6 and 7. As each principal component is individually normalized, an amplitude of more than 2.8σ is unusual at the 5% level. (b) Amplitude of the wave 2 component of the 60°N 500 hPa geopotential height (thick curve). The projection of wave 2 at 150°W (North Pacific) and at 60°E (Eurasia) are shown by the thin and dashed curves respectively. (c) Strength of the SPV measured by the zonal mean zonal wind (m s⁻¹) at 60°N and 10 hPa. (d) NAO index (hPa), taken as the absolute PMSL difference between the Azores and Iceland. Date intervals are weeks (markers correspond to Mondays) and the dashed grey line indicates the central date of the SSW.
reproduce the SSW. As highlighted above, its North Pacific response is too weak and it possesses little of the Eurasian signal. This means its wave 2 forcing is too small to consistently trigger an SSW in its members.

### 3.4 | January 2019 SSW

Weakening of the SPV started in mid-December 2018, leading to a wave 1 driven SSW on the 1st January (Lee and Butler, 2020). GloSea5 gave a signal for a weak SPV, particularly in January and February, even from November (Figure 5). This is likely to be related to the weak El Niño underway at the time. Indeed, tropical conditions appear to have increased the chances of this outcome even in December. Unlike the previous winter’s event, downward propagation of observed easterly wind anomalies appears to stall at the tropopause, with westerly wind anomalies in the troposphere during most of January and easterly anomalies appearing in the troposphere only at the start of February. Despite this, the stratospheric relaxation experiment shows the SSW’s impact on the troposphere is to produce easterly anomalies. The duration and intensity of the SSW and the tropospheric response is at least as large as for the 2018 SSW (Figure 2). The observed differences in the troposphere therefore result from processes unrelated to the stratosphere. For part of January, at least, there is a westerly influence on the zonal wind from the Tropics, in opposition to the influence from the stratosphere. This contending factor, therefore, can explain why tropospheric easterlies were masked and delayed in 2019. A similar masking occurred later in March 2018 after the main easterly event (Figure 2). Following the 2019 SSW, the SPV recovered by mid-February, becoming stronger than usual, as often occurs after mid-winter events (Limpasuvan et al., 2004). The recovery signal is seen to an extent in the tropical forcing experiment, consistent with the signal for a weaker SPV earlier in winter.

Monthly PMSL patterns in winter 2018–19 are shown in Figure 6. GloSea5 predicted December’s pattern of North Atlantic-European anomalies moderately well ($r = 0.32$), showing a positive NAO-like signal tilted south west to north east (with an NAO index anomaly of +2.6 hPa compared to +3.2 hPa in ERA Interim). In addition, it predicted some of the broad high-latitude anomalies. The duration and intensity of the SSW and the tropospheric response is at least as large as for the 2018 SSW (Figure 2). The observed differences in the troposphere therefore result from processes unrelated to the stratosphere. For part of January, at least, there is a westerly influence on the zonal wind from the Tropics, in opposition to the influence from the stratosphere. This contending factor, therefore, can explain why tropospheric easterlies were masked and delayed in 2019. A similar masking occurred later in March 2018 after the main easterly event (Figure 2). Following the 2019 SSW, the SPV recovered by mid-February, becoming stronger than usual, as often occurs after mid-winter events (Limpasuvan et al., 2004). The recovery signal is seen to an extent in the tropical forcing experiment, consistent with the signal for a weaker SPV earlier in winter.

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![Figure 5](image-url)  
**Figure 5** As Figure 2, but for winter-spring 2018–19
negative AO pattern in January that was consistent with the response to the emerging SSW (Figure 6, top middle). In mid-latitudes, however, the low-pressure anomalies expected over central and southern Europe were only realized over the eastern part of the continent, and the North Atlantic patterns were not predicted \((r = -0.34)\). The tropical relaxation experiments reproduce the anticyclonic anomaly that was observed in the central mid-latitude Atlantic \((r = 0.61)\). This feature represents a positive phase of the East Atlantic (EA) pattern, the second most important mode of North Atlantic variability after the NAO (Barnston and Livezey, 1987). It counteracted the low-pressure anomalies produced by the stratosphere ensuring milder, northwesterly flow into Western Europe, rather than easterly flow. In February, the North Atlantic PMSL reverts to a pattern similar to that seen in December, with a low-pressure anomaly in the mid-Atlantic replacing the high-pressure anomaly. Both tropical and stratospheric influences align to give negative NAO-like patterns, suggesting that February 2019 was a candidate to be a cold month in Europe. Indeed, there were zonal mean easterly anomalies in this month, but a high-pressure anomaly over central and southern Europe introduced mild, southwesterly flow. This last anomaly is not produced in either the relaxation experiments \((r = 0.26\) and \(-0.11\) in the tropical and stratospheric experiments respectively), suggesting that it may be a result of internal variability generated by mid-latitude
dynamics. Consequently, 2019 represents a ‘near miss’ for an extreme cold spell. In March, and later in spring, the influence of the stratosphere becomes weaker, with North Atlantic-European anomalies mainly resulting from tropical forcing (not shown).

4 CONCLUSIONS

Ensemble predictions from GloSea5 showed signals for positive winter-mean NAO in 2017–18, and negative NAO in 2018–19, related to the shift from weak La Niña to weak El Niño conditions. A major SSW in February 2018 occurred too late to substantially change winter-mean temperatures but produced impactful cold spells across much of Europe. In contrast, observed effects of the January 2019 SSW were limited. Numerical relaxation experiments isolating the remote forcing of the mid-latitude atmospheric circulation show that tropical influences had opposite effects in the two winters. In 2018, they were crucial for the initiation of the SSW and enhanced its response via tropospheric teleconnections. In 2019, however, tropical forcing led to an Atlantic pressure pattern that prevented the development of cold weather in Europe. Rao et al., 2020, using sub-seasonal forecasts, suggest that the 2018 SSW had greater potential to affect the surface than the 2019 SSW. These results may relate to predictability, however, rather than highlighting the stratosphere-troposphere connections. Here, using experiments that are designed to isolate this aspect, we show that both SSW events actively influenced the troposphere, but in one case a conflicting influence overcame the stratospheric effect. Apparent differences in SSW responses more generally may simply be due to competing influences such as those from the Tropics or internal variability (Gerber et al., 2009).

The tropospheric wave 2 forcing that gave rise to the February 2018 SSW is related to an extreme MJO event. This event, and increasing chances of weakening of the SPV, were predicted by GloSea5 from late January. The Pacific anticyclonic node of the wave 2 pattern is a response to MJO phases 6/7. Its Eurasian counterpart, which enhanced wave 2 only for a short period, appears to arise from internal variability or forcings other than the Tropics or stratosphere. Although sub-seasonal forecasts initialized about 10 days before the SSW (Lee et al., 2019) have been used to suggest an Atlantic synoptic trigger, Pacific signals are likely to be already well-represented in these forecasts, masking the importance of the MJO event. The findings of the current study illustrate the benefit of revisiting recent seasons and their forecasts to obtain deeper understanding of mechanisms and build confidence in predictions.

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