

1 **The Downward Influence of Sudden Stratospheric Warmings: Association**
2 **with Tropospheric Precursors**

3 Ian White*

4 *The Hebrew University of Jerusalem, Institute of Earth Sciences, Edmond J. Safra Campus, Givat*
5 *Ram, Jerusalem, Israel*

6 Chaim I. Garfinkel

7 *The Hebrew University of Jerusalem, Institute of Earth Sciences, Edmond J. Safra Campus, Givat*
8 *Ram, Jerusalem, Israel*

9 Edwin P. Gerber

10 *Courant Institute of Mathematical Sciences, New York University, New York, USA*

11 Martin Jucker

12 *Climate Change Research Centre, University of New South Wales, Parkville, Sydney, Australia*

13 Valentina Aquila

14 *American University, Dept of Environmental Science, Washington, DC, USA*

15 Luke D. Oman

16 *NASA Goddard Space Flight Center, Greenbelt, Maryland, USA*

¹⁷ **Corresponding author address:* Ian White, The Hebrew University of Jerusalem, Institute of Earth
¹⁸ Sciences, Edmond J. Safra Campus, Givat Ram, Jerusalem, Israel.
¹⁹ E-mail: ian.white@mail.huji.ac.il

ABSTRACT

20 Tropospheric features preceding Sudden Stratospheric Warming events
21 (SSWs) are identified using a large compendium of events obtained from a
22 chemistry-climate model. In agreement with recent observational studies,
23 it is found that approximately one third of SSWs are preceded by extreme
24 episodes of wave activity in the lower troposphere. The relationship becomes
25 stronger in the lower stratosphere, where $\sim 60\%$ of SSWs are preceded by ex-
26 treme wave activity at 100 hPa. Additional analysis characterises events that
27 do or do not appear to subsequently impact the troposphere, referred to as
28 downward and non-downward propagating SSWs, respectively. On average,
29 tropospheric wave activity is larger preceding downward-propagating SSWs
30 compared to non-downward propagating events, and associated in particular
31 with a doubly-strengthened Siberian High. Of the SSWs that were preceded
32 by extreme lower-tropospheric wave activity, $\sim 2/3$ propagated down to the
33 troposphere, and hence the presence of extreme lower-tropospheric wave ac-
34 tivity can only be used probabilistically to predict a slight increase or decrease
35 at the onset, of the likelihood of tropospheric impacts to follow. However,
36 a large number of downward and non-downward propagating SSWs must be
37 considered (> 35), before the difference becomes statistically significant. The
38 precursors are also robust upon comparison with composites consisting of
39 randomly-selected tropospheric NAM events. The downward influence and
40 precursors to split and displacement events are also examined. It is found that
41 anomalous upward wave-1 fluxes precede both cases. Splits exhibit a near
42 instantaneous, barotropic response in the stratosphere and troposphere, while
43 displacements have a stronger long-term influence.

44 **1. Introduction**

45 Approximately once every other year, the winter-hemisphere westerly stratospheric Polar
46 Vortex weakens, reverses in direction and warms dramatically over the course of just a few days
47 in a sudden stratospheric warming (hereafter SSW; see Butler et al. 2015, and references therein).
48 Generally it is thought that such a SSW is caused by an anomalously strong upward flux of
49 planetary waves from the troposphere (e.g., Matsuno 1971; Polvani and Waugh 2004; Sjoberg
50 and Birner 2012). However, it is not known if the reason for this upward flux into the stratosphere
51 is due to an anomalously large generation of wave activity in the troposphere, or due to the
52 stratosphere being in such a state as to take advantage of the large reservoir of tropospheric wave
53 activity and encourage anomalous wave propagation through the tropopause (Jucker 2016; Birner
54 and Albers 2017; de la Camara et al. 2017). Due to the hemispherical differences in topography,
55 all but one of the observed SSWs have occurred in the Northern hemisphere (NH) (e.g., Charlton
56 and Polvani 2007).

57
58 It is acknowledged that SSWs can have an appreciable influence on the tropospheric circulation
59 below for up to 2 months following the onset of the event (e.g., Baldwin and Dunkerton 2001;
60 Nakagawa and Yamazaki 2006; Mitchell et al. 2013; Hitchcock and Simpson 2014; Kidston et al.
61 2015). In particular, SSWs on average precede a persistent equatorward shift of the North Atlantic
62 eddy-driven jet (i.e., a negative phase of the North Atlantic Oscillation [NAO]). The eddy-driven
63 jet is colocated with the extratropical storm tracks, and hence plays a crucial role in determining
64 the weather over North America and Europe (e.g., Kidston et al. 2015). Additionally, it has been
65 shown that SSWs result in an increase in cold-air outbreaks in the midlatitude NH (Thompson
66 et al. 2002; Tomassini et al. 2012) as well as high-latitude blocking events (Martius et al. 2009).

67 Thus, it has been suggested that the skill of tropospheric seasonal forecasts can be improved
68 by enhancing our understanding of SSWs and their downward influence on the tropospheric
69 circulation (Marshall and Scaife 2010; Scaife et al. 2012; Smith et al. 2012; Sigmond et al. 2013;
70 Tripathi et al. 2014).

71
72 Whilst there is a clear aggregate impact of SSWs on the troposphere, there is considerable
73 variation between individual events (Baldwin and Dunkerton 2001; Sigmond et al. 2013). Indeed,
74 some events exhibit no visible impact and hence this has led to studies defining SSWs as either
75 'downward' (DW) or 'nondownward' (NDW) propagating (Jucker 2016; Kodera et al. 2016;
76 Runde et al. 2016; Karpechko et al. 2017). However, there is debate about whether there is
77 an actual DW communication of information from the stratosphere, or whether the observed
78 influence is related to variability inherent to the troposphere (Kidston et al. 2015).

79
80 Previous studies have highlighted the role of the stratosphere in determining the extent of
81 the DW influence. It has been suggested that the type and magnitude of the wave forcing (be
82 it wave-1 or wave-2) entering the stratosphere (e.g., Nakagawa and Yamazaki 2006), the type
83 of SSW (split or displacement) which occurs (e.g., Mitchell et al. 2013; Seviour et al. 2013;
84 O'Callaghan et al. 2014; Seviour et al. 2016), the depth to which the initial warming descends in
85 the stratosphere (Gerber et al. 2009; Hitchcock et al. 2013), and the persistence of the SSW in the
86 lower stratosphere (Hitchcock and Simpson 2014; Maycock and Hitchcock 2015) can all play a
87 role, either individually or collectively, in determining the tropospheric response. For instance,
88 Nakagawa and Yamazaki (2006) found that observed SSW events which were followed by a
89 significant long-lasting tropospheric anomaly were associated with an enhanced upward flux of
90 wave 2. Mitchell et al. (2013) and Seviour et al. (2013) found that the observed tropospheric

91 response was dependent on the SSW type; split SSWs were associated with such a response,
92 whereas displacement SSWs were not. Recently, using a large compendium of modelled SSWs,
93 Maycock and Hitchcock (2015) found only small differences between both types, but also
94 that the surface responses were not robust to the algorithm used to classify the events. They
95 also suggested that the tropospheric impact was dependent on whether the lower-stratospheric
96 circulation anomalies persisted; a point which was also proposed by Hitchcock and Simpson
97 (2014) and Karpechko et al. (2017) using reanalysis data and a full chemistry-climate model, as
98 well as by Jucker (2016) using idealised GCM experiments. Lehtonen and Karpechko (2016) and
99 Karpechko et al. (2017) both indicated the role of enhanced upward-propagating planetary waves
100 prior to the onset of the SSW as well as its continuation for a up to a week after the onset.

101

102 On the other hand, both observational and modelling studies have suggested that the troposphere
103 may play a role in the initial forcing of some SSW events (e.g., Martius et al. 2009; Garfinkel
104 et al. 2010; Cohen and Jones 2011; Dai and Tan 2016; Hitchcock and Haynes 2016; Bao et al.
105 2017) as well as the ensuing tropospheric response be it due to the state of the troposphere prior
106 to the onset (Black and McDaniel 2004) or due to the presence of synoptic-scale eddy feedbacks
107 (Limpasuvan et al. 2004; Song and Robinson 2004; Domeisen et al. 2013; Hitchcock and Simpson
108 2014). However, whilst precursors such as blocking events have been found to occur before 25
109 of the 27 SSWs observed in ERA-40 (Martius et al. 2009), only 6% of blocking events during
110 1957-2001 were actually followed by a SSW. These results indicate that tropospheric precursors
111 are perhaps not a useful predictor, despite them occurring prior to many SSWs. Garfinkel et al.
112 (2010) found that surface variability over the North Pacific and Eastern Europe could either
113 deepen or flatten the troughs/ridges associated with tropospheric stationary planetary waves. Such
114 precursors over these two regions then lead to changes in the upward wave flux and possibly the

115 onset of a weaker Polar Vortex, followed by its DW propagation. Depending on the magnitude
116 and spatial location of this anomalous forcing, either a split or displacement SSW may occur (e.g.,
117 Cohen and Jones 2011). Further, Black and McDaniel (2004) observed that the determination
118 of the DW propagation of a SSW depended on the pre-existing tropospheric state; in the case of
119 nondownward-(NDW)-propagating events, the troposphere was already in a positive NAM-like
120 state which acted to mask the DW stratospheric influence. In the case of DW-propagating
121 events, the troposphere was already in a negative NAM-like state, although slightly out of phase,
122 latitudinally, with the canonical NAM.

123

124 In contrast, modelling studies by Gerber et al. (2009) and Hitchcock and Simpson (2014)
125 suggest that differences between DW and NDW events are associated primarily with differences
126 in tropospheric variability. That is to say, they hypothesize that there is a deterministic influence
127 of SSWs on the troposphere (a forced response), which is combined with an essentially stochastic
128 component associated with internal tropospheric variability. The latter can mask/enhance the DW
129 forced signal and thus predicting the response to a SSW will likely be limited by our ability to
130 forecast tropospheric weather. This also speaks to the difficulty in being able to understand the
131 mechanisms behind the DW propagation of a SSW.

132

133 One of the key aims of this paper is to identify and determine the robustness of tropospheric
134 precursory features to SSWs as well as to assess whether these tropospheric precursors may be
135 important for discriminating between DW and NDW SSWs, using a large compendium of SSWs
136 obtained from the Goddard Earth Observing System Community Climate Model (GEOSCCM).
137 The paper then has the following structure: in section 2 we present a description of the GEOSCCM
138 model integrations used in this study, and of the methods used to identify SSWs (Charlton and

139 Polvani 2007) and split and displacement vortex events (Seviour et al. 2013), and also determine
140 whether these events are DW or NDW propagating (Jucker 2016; Runde et al. 2016; Karpechko
141 et al. 2017); in section 3 we present the results; and finally, in section 4 we present a summary
142 and discussion.

143

144 **2. Methodology**

145 *a. Model Output*

146 We utilise a series of model integrations which were performed using the Goddard Earth
147 Observing System Chemistry-Climate Model, Version 2 (GEOSCCM; see Rienecker et al. 2008).
148 The GEOSCCM couples the GEOS-5 (Molod et al. 2012) atmospheric general circulation model
149 (GCM) with StratChem, a comprehensive stratospheric chemistry module (Pawson et al. 2008).
150 In total, 40 historical-run integrations are here analysed, 25 of which are of length 30 years
151 (January 1980 to December 2009) and 15 are of length 55 years (January 1960 to December
152 2014), which yields a total of 1575 years of data to analyse. These are described in more detail
153 in Garfinkel et al. (2015), Aquila et al. (2016) and Garfinkel et al. (2017). The integrations were
154 performed for different purposes and therefore this ‘super ensemble’ encompasses a range of
155 forcings and physical parameterisations. These include changing sea surface temperatures, sea-ice
156 and greenhouse gas concentrations, as well as ozone-depleting substances, solar variability,
157 and volcanic eruptions. We note that there is a slight influence of SSTs on the DW and NDW
158 propagation of SSWs with there being slightly more DW SSWs than NDW SSWs during El Nino
159 years, but it is comparatively weak and this will be discussed in a future publication. We also note
160 that the two different time periods (i.e., pre- and post-satellite era) over which the integrations

161 are run do not have an influence on the results. The model was run using 72 vertical layers with
162 a lid at 0.01 hPa, although we base our analysis on 14 levels ranging from 700 hPa up to 1 hPa.
163 We note that at 700hPa, there were small areas over mountain regions for which no value was
164 outputted from the model; these were filled in using an interpolation scheme in this study so that
165 we could decompose the heat flux into different zonal wavenumbers. The horizontal resolution is
166 2° latitude by 2.5° longitude.

167

168 *b. SSW Definitions*

169 To define SSW events in the GEOSCCM model integrations described above, we first utilise
170 a simplified version of the World Meteorological Organisation (WMO) criteria proposed by
171 Charlton and Polvani (2007) where SSWs are defined by a reversal of the zonal-mean zonal
172 wind \bar{u} at 60°N and 10 hPa to easterly winds from November 1st to March 31st. This criterion
173 is supplemented by the requirement that winds return to a westerly state for a period of 10
174 consecutive days prior to April 30th, which helps avoid counting any final warmings, and a
175 separation of at least 20 days between two consecutive events, to avoid counting the same SSW
176 event twice (see also the corrigendum of Charlton and Polvani 2007). Using the SSW definition
177 above, a total of 962 SSWs (see table 1) are found giving a ratio of 0.61 per year; a ratio which
178 is a little smaller than that found in observations (also see table 1 in Butler et al. 2015). We note
179 that this slight decrease in the SSW frequency relative to that observed may be due to the fact that
180 the climatological planetary-wave flux entering the stratosphere near 100 hPa in our 40 runs is
181 smaller than in ERA-Interim.

182

183 We also identify the two characteristic types of extreme vortex variability - split and displace-
184 ment SSWs - using the 2-D moment analysis method described by Seviour et al. (2013). In
185 particular, the geopotential height Z at 10 hPa, rather than the potential vorticity as in Mitchell
186 et al. (2013), is used in this method. Seviour et al. (2013) detail this method, but there are three
187 parameters which are modified for this study. The first is the edge of the Polar Vortex, which we
188 here define as the December-March (DJFM) climatological mean Z at 60°N and 10 hPa (as in
189 Maycock and Hitchcock 2015), where the climatology is defined as the average during DJFM in
190 all 40 ensemble members. The second and third are the thresholds for the split and displacement
191 SSWs, which depend on the values of the centroid latitude and aspect ratio. We here choose the
192 thresholds as the most equatorward 5% of centroid latitudes and largest 5% of aspect ratios in
193 all ensemble members, yielding thresholds of 64.38°N and 2.074 respectively (compare these
194 values to the respective 5.7%/66°N and 5.2%/2.4 used in Seviour et al. 2013). We note that the
195 results are not sensitive to slight changes in the thresholds used here. We also note that a handful
196 of events satisfy both criteria, in which case they are marked as unclassifiable, to try and best
197 ensure independent events. Using this method, we find a total of 903 events with 400 splits, 500
198 displacements, and 3 unclassified (see table 1). Note that these events are not the same as the 962
199 SSW events identified using the CP07 method, as we do not here classify the CP07-identified
200 SSWs as splits or displacements. Nevertheless, 545 of the CP07-identified SSWs overlap within
201 ± 10 days of an identified displacement or split SSW.

202

203 *c. DW- and NDW-propagating event definitions*

204 To define whether a given event is DW or NDW propagating we utilise the NAM index. In
205 this study we compute a simplified NAM index based on the polar-cap average geopotential

206 height, Z (Baldwin and Thompson 2009). Standardised Z anomalies are calculated at each level
207 as the deviation from the 60-day low-pass filtered daily climatology, which are subsequently
208 smoothed using a 3-day running mean, following Martineau and Son (2015), although we note
209 that quantitatively similar results can be found using different filtering windows. The anomalies
210 are then area-averaged (i.e., multiplied by $\cos \varphi$ where φ is latitude) over $60\text{-}87^\circ\text{N}$, divided by
211 the standard deviation at each level and multiplied by -1 so that conventionally, a negative NAM
212 index identifies with a positive Z anomaly and vice versa.

213

214 Four definitions have been proposed recently to characterise the DW propagation of SSWs
215 using the NAM index; one by Runde et al. (2016), two by Jucker (2016), and one by Karpechko
216 et al. (2017). In this manuscript we mostly present results using that by Karpechko et al. (2017)
217 and hence this is the one we briefly summarise here. The descriptions of the other three are
218 included in the supplementary material. Karpechko et al. (2017) introduced three criteria that
219 must be satisfied, these being: 1) the averaged NAM index at 1000 hPa over the period ranging
220 from 8 days until 52 days after the onset date must be negative; 2) the fraction of days in this
221 45-day period on which the NAM index at 1000 hPa is negative must be greater than 0.5; and 3)
222 the fraction of days in this 45-day period on which the NAM index at 150 hPa is negative must
223 be greater than 0.7. Note that for the first two criteria we use the NAM at 850 hPa to reduce
224 complications with topography and for the third we use 100 hPa to ensure that the anomalies
225 persist in the lower stratosphere, although we note that the results are not sensitive to the choice
226 of level. These criteria are chosen to ensure that there is a long-lasting tropospheric signal of
227 the negative NAM anomalies associated with the upper-tropospheric/lower-stratospheric negative
228 anomalies. See table 1 for the numbers of DW and NDW SSWs resulting from all four DW

229 definitions.

230

231 **3. Results**

232 We start by identifying apparent precursory features to SSWs (both DW- and NDW-propagating)
233 using composites over all of the modelled SSW events. We then test the robustness of these
234 precursors using different DW definitions as introduced in section 2 and random composites of
235 tropospheric events, before examining the number of SSWs which are actually preceded by these
236 precursors. Finally, we briefly examine the precursory features to splits and displacements along
237 with their division into DW and NDW events. Note that herein we define a precursor to be an
238 anomalous feature which is found to occur prior to a SSW event, but do not claim there to be
239 any deterministic aspect, as there is no one-to-one relationship between any of the precursors
240 we identify and the subsequent stratospheric state due to the large internal variability of the
241 stratosphere.

242

243 *a. Composite Analyses of DW and NDW Events*

244 As a starting point, we examine the evolution of the NAM index which has been traditionally
245 used as a measure of stratosphere-troposphere coupling. The NAM for all SSWs is composited
246 at lag zero according to the onset date of the SSW (see section 2). We only show results using
247 the DW definition of Karpechko et al. (2017) but note that the robustness of these results to DW
248 definition is discussed in section 3b. Figure 1 shows the NAM index composited over a) all
249 SSW events in all of the ensemble members (a total of 962; see table 1); b) all DW-propagating
250 SSW events (506; as determined by the criteria in Section 2); c) all NDW-propagating SSW

251 events (456); and d) the composite difference between the DW- and NDW-propagating events
252 (hereafter DW-NDW). In the all event composite (a), the NAM index is similar to the canonical
253 'dripping-paint' pattern first highlighted by Baldwin and Dunkerton (2001). The negative
254 anomalies initialise around lags -15 to -10 above ~ 250 hPa, and at lag zero maximise in the
255 upper stratosphere. The negative anomalies propagate DW to the lower stratosphere over the
256 next few weeks and start to recover in the upper stratosphere after lag +20, although those in the
257 lower stratosphere persist until lag +60. Negative anomalies are visible in the troposphere for all
258 positive lags, but with much smaller amplitude than those in the stratosphere.

259

260 Upon subdividing the total into DW- and NDW-propagating events (b and c), it can be seen that
261 the DW events have a much stronger influence on the troposphere after lag 0, by construction,
262 with negative NAM anomalies reaching down to near the surface and persisting for over 60 days.
263 At positive lags, the DW composite (b) has magnitudes of around twice that of the total (a) in
264 the troposphere, which is due to the cancellation between the negative DW anomalies and the
265 weakly-positive NDW anomalies in (c). Further, the magnitude of the negative anomalies in the
266 upper stratosphere is larger for the DW events, and those in the lower stratosphere persist for
267 considerably longer during DW events. Finally, there are larger negative tropospheric anomalies
268 in the DW composite compared to the NDW composite prior to lag zero. Zonal-mean anomalies
269 prior to lag zero have been found with both the same sign (Jucker 2016; Karpechko et al. 2017)
270 and also with opposite sign (Hitchcock and Haynes 2016) using a large compendium of modelled
271 SSWs. To this point, Gerber et al. (2010) showed such precursor anomalies to be both model-,
272 as well as configuration-dependent. For instance, Gerber et al. (2010), using the Canadian
273 Middle Atmosphere Model (CMAM) found such precursors, but using a slightly different model
274 configuration, Hitchcock and Simpson (2014) did not. It appears that DW SSW events appear to

275 be stronger in overall magnitude in both the troposphere and stratosphere, persist for longer in
 276 the lower stratosphere and have evidence of tropospheric preconditioning, in comparison to those
 277 which are NDW propagating.

278

279 To examine the differences in upward wave activity between DW and NDW events, in figure 2
 280 we show the height-time evolution of the vertical component of the Eliassen-Palm (EP) flux

$$F^{(z)} = \rho_0 a \cos \varphi \left(\left[f - \frac{1}{a \cos \varphi} (\bar{u} \cos \varphi) \right] \overline{v' \theta' / \theta_z} - \overline{w' u'} \right) \quad (1)$$

281 (Andrews and McIntyre 1978; Andrews et al. 1987), where φ and z are the latitude and
 282 log-pressure height coordinates, u , v and w are the zonal, meridional and vertical components
 283 of the wind, θ is the potential temperature, f , a and ρ_0 are the Coriolis parameter, Earth's
 284 radius and basic-state density, and overbars and primes represent the zonal-mean and deviations
 285 from the zonal-mean, respectively. $F^{(z)}$ is averaged over the latitude band of 45-75°N and
 286 filtered for planetary waves 1 and 2, and as in figure 1, presented as composites over (a) all
 287 SSWs, (b) DW SSWs, (c) NDW SSWs, and (d) the DW-NDW difference. As advocated by
 288 Jucker (2016) and Birner and Albers (2017), the anomalies are standardised by dividing each
 289 level by the climatological standard deviation so that, for example, a value of 2 represents a 2
 290 standard-deviation from the mean. This allows one to determine how strong the wave bursts at
 291 a given level are, compared to general variability at that level (Jucker 2016; Birner and Albers
 292 2017). Prior to the onset date, it is clear that in the all, DW and NDW composites, the anomalous
 293 wave flux at stratospheric levels is in a relative sense, larger than at tropospheric levels. In
 294 particular, in the DW composite, the anomalies have a magnitude of nearly 2.5 standard deviations
 295 in the stratosphere and of 0.75 standard deviation in the troposphere, whereas in the NDW
 296 composite, the values are comparatively small with values of 2 and 0.25 standard deviations in

297 the stratosphere and troposphere. The gradual upward propagation at negative (-30 to -15) lags
298 also hints that for some events, there is a tropospheric source of wave activity which may well be
299 amplified in the stratosphere closer to the onset date. The DW-NDW composite makes clearer the
300 significant differences with values of around 0.25-0.5 standard deviations, becoming largest in the
301 stratosphere closer to the onset date.

302
303 At positive lags, the anomalies in both the DW and NDW composites are negative in the
304 stratosphere indicating reduced upward wave propagation after the onset date. However, we note
305 that the positive anomalies around the onset date do persist in the stratosphere for up to a week.
306 In the troposphere, the anomalies are of opposite sign between DW and NDW events; for the
307 DW events, there are weakly positive anomalies (in this standardised sense - if using the full field
308 then they become larger) which we note are dominated by wave-2, whereas for NDW events,
309 there are negative anomalies. The weakly positive anomalies for DW events are seemingly in
310 disagreement with Hitchcock and Simpson (2014) and Hitchcock and Haynes (2016) who found
311 reduced vertical wave flux during the recovery phase, but since they are of very small magnitude
312 compared to tropospheric variability, we don't expect the difference between this feature and
313 the aforementioned studies to be significant. We also note that synoptic waves contribute in the
314 troposphere at positive lags (not shown).

315
316 These $F^{(z)}$ anomalies allow us to define certain lag stages in the evolution of the DW and
317 NDW SSWs (see dashed vertical lines). The first is the preconditioning stage (hereafter PC)
318 from lags -25 to -1, and these lags are chosen as they represent the approximate duration of the
319 significant tropospheric precursor DW-NDW differences, although we note that the tropospheric
320 and stratospheric anomalies intensify at around lag -15. The second is the onset stage (ONS) from

321 lags 0 to +5, which is associated with continued (reduced) anomalous upward wave propagation in
322 the stratosphere (troposphere). Finally, we classify the recovery stage (REC) over lags +6 to +50
323 which represents the approximate timescale over which the tropospheric DW-NDW differences
324 disappear. Note that results in this paper are not sensitive to slight changes in the definition of
325 these lags.

326
327 With the zonal-mean NAM precursors in mind (figure 1), we now determine if there are any
328 such precursors in a latitude-longitude sense. In figure 3 we show Z anomalies at 700 hPa
329 averaged over the PC stage (top row), ONS stage (middle row), and REC stage (bottom row).
330 The November-February climatology for each variable is superimposed as green contours and we
331 note that the climatologies in these GEOSCCM integrations agrees well with observations (e.g.,
332 Garfinkel et al. 2010).

333
334 In the PC stage, the Z anomalies for the DW (a) and NDW (b) composites show similar
335 spatial patterns, with a clear wave-1 like structure consisting of negative anomalies northward
336 of 60°N over the North Pacific and positive anomalies over Scandinavia and Europe. These
337 negative (positive) anomalies project onto the climatological stationary planetary wave-1 centres
338 of action, albeit slightly offset to the northeast (northwest), respectively. In the DW composite,
339 the magnitudes of the anomalies are noticeably larger than in the NDW composite; in particular
340 the positive anomalies over Northern Europe are doubled in the DW composite. This difference
341 in magnitudes is highlighted in the DW-NDW composite (top right) with negative and positive
342 differences over the Aleutian Low sector and the Siberian High sector respectively. We also
343 note the regions of positive and negative anomalies further equatorward over the North Pacific
344 and North Atlantic respectively. Over the North Atlantic, the anomalies are significantly more

345 negative for the DW events.

346

347 During the ONS stage (middle row), positive anomalies appear over the Polar cap with an
348 annulus of negative anomalies starting to develop at midlatitudes for the DW events. For the NDW
349 events however, positive and negative anomalies develop over the Aleutian Low and Siberian
350 High regions, respectively, projecting negatively onto the climatological centres and suggesting
351 a reduced upward wave-1 flux. This yields differences which still show a wave-1 pattern over
352 the North Pacific and Siberia, along with more widespread negative differences over the North
353 Atlantic (compared to during the PC stage). The latter highlights the canonical DW influence
354 of SSWs. The NAM at lags 0 to +5 is not utilised in the Karpechko et al. (2017) DW definition
355 and hence these anomalies are not forced by the averaging associated with the definition. During
356 the REC stage (bottom row), the strongest anomalies are associated with the DW events (indeed,
357 with much smaller anomalies in the NDW composite), which exhibit a highly zonal pattern,
358 with positive anomalies at high latitudes surrounded by an annulus of negative anomalies at
359 midlatitudes, projecting onto the negative phase of the NAO. While the annulus pattern during
360 REC is present by construction, the DW-NDW difference during the PC and ONS stages are not.

361

362 In the previous three figures, there is clearly on average, enhanced upward wave activity in the
363 troposphere, a more negative tropospheric NAM and an enhanced Siberian High for DW events
364 prior to the SSW onset. We now further examine the connection between these three features in
365 figure 4, but, instead of splitting the SSWs according to the sign and magnitude of the NAM after
366 the onset (as in figures 1- 3), we split them according to the strength of $F^{(z)}$ (filtered for waves
367 1-2) in the lower troposphere, before the onset date. In particular, we composite the SSWs into (a;
368 d; g) the half of SSWs with the smallest $F^{(z)}$ at 500 hPa, averaged over lags -15 to -1 (SSW_{small}),

369 and (b; e; h) the half of SSWs with the largest such $F^{(z)}$ (SSW_{large}). In (c; f; i), the SSW_{large} -
370 SSW_{small} differences are then shown. In the top row, the clear feature is the larger $F^{(z)}$ anomalies
371 throughout the troposphere and stratosphere at negative lags in SSW_{large} events, although note
372 that the lower-tropospheric anomalies at negative lags are by construction.

373

374 In the middle row (the NAM index), it is clear that the tropospheric NAM is more negative
375 for SSW_{large} events at both negative and positive lags as well as being more negative in the
376 stratosphere after the onset. Finally, in the bottom row (Z), the clearest differences between the
377 SSW_{large} and SSW_{small} events are the negative and positive anomalies over the North Pacific and
378 Siberian High regions, respectively, which are much enhanced for the SSW_{large} events. These
379 project positively onto the climatological centres of action, and thus are likely linked with the
380 enhanced $F^{(z)}$ seen in the top row. Together with the $F^{(z)}$ panels, the NAM and Z anomalies
381 suggest that enhanced upward lower-tropospheric wave activity prior to the SSW onset date may
382 lead to a weaker Polar Vortex and subsequently be associated with a more negative tropospheric
383 NAM after the onset.

384

385 In order to determine the vertical extent of the Z anomalies, we show longitude-height
386 cross-sections of Z' (i.e., the deviation from zonal-mean) in figure 5, averaged over the same lag
387 stages as in figure 3 and over the latitude band of 50-60°N. This latitude band is chosen as it best
388 captures the negative and positive anomalies over the Aleutian Low and Siberian High regions
389 shown in figure 3. In the climatology (thin black contours), there is a clear westward tilt with
390 height of Z' agreeing with the well-known westward tilt of upward-propagating planetary waves
391 (e.g., Andrews et al. 1987). The Z' has a wave-1 structure in the stratosphere with one ridge and
392 one trough, but is associated with higher wavenumbers in the troposphere (multiple ridges and

393 troughs). This agrees with the Charney-Drazin criterion (Charney and Drazin 1961) which states
394 that only planetary waves can propagate into the stratosphere and smaller-scale waves are limited
395 to propagation in the troposphere.

396
397 During the PC stage (top row), the anomalies for both DW and NDW events project posi-
398 tively onto the climatological Z' anomalies and exhibit the canonical westward tilt with height,
399 indicating anomalous upward wave propagation from the troposphere to the lower-to-middle
400 stratosphere. In particular, in the troposphere, there are negative anomalies spanning from 70°E
401 eastward to ~150°W, and positive anomalies from 150°W eastward to ~60°E. These agree with
402 the Z' anomalies at 700 hPa shown in figure 3. In the difference plot, it is clear that the anomalies
403 associated with DW events are generally larger in magnitude indicating enhanced upward wave
404 propagation.

405
406 After the onset date (middle row), the anomalies above 10 hPa change sign, thus projecting
407 negatively onto the climatological centres. This is likely associated with reduced upward wave
408 propagation deep into the stratosphere after a SSW event, in agreement with the Charney-Drazin
409 criterion. Below 50 hPa, the anomalies and differences look generally similar to during the PC
410 stage although slightly more connected, suggesting continued upward wave propagation into
411 the lower stratosphere. During the REC stage (bottom row), the upper-to-middle stratospheric
412 anomalies extend deeper into the lower stratosphere compared to during the ONS stage and are
413 still of opposite sign to the climatology. The latter point indicates that waves are absent above 50
414 hPa under DW events, and much reduced under NDW events. This is in agreement with a SSW
415 event which has a more negative NAM (figure 1). Below 50 hPa, they lose their westward tilt with
416 height, instead either exhibiting more of an eastward tilt, particularly over the North Pacific (g),

417 or vanishing almost entirely (h).

418

419 It is worthwhile to examine how many SSWs are required to find precursory features such as
420 those found in figures 1- 5. For instance, these precursor features to DW and NDW events are
421 not found in reanalysis products such as the ERA-Interim reanalysis (see figure 1 in Karpechko
422 et al. 2017), but they have been found in large-samples obtained from GCMs (e.g., figure 3 in
423 Karpechko et al. 2017). Hence in figure 6 we plot confidence intervals of the DW-NDW difference
424 for the PC stage (-25 to -1) of (a) the NAM index at 700 hPa, (b) $F^{(z)}$ at 700 hPa averaged over
425 45-75°N, and (c) Z at 700 hPa area averaged over 50-80°N, 60-90°E, i.e., the positive differences
426 slightly northwest of the climatological Siberian High. The confidence intervals are estimated
427 using a Monte-Carlo repeat sampling procedure (100,000 repetitions), for different prescribed
428 sample sizes. The confidence intervals for the 90% (red), 95% (green) and 99% (blue) levels all
429 converge to the overall composite mean shown in the corresponding figures (see dotted black
430 lines), as the sample size is increased from the minimum of 10 considered here, to the maximum
431 of 455. From the definition of a confidence interval around the difference between the means of
432 two samples, if the interval does not contain zero, then the means are significantly different from
433 one another, at the chosen level. Hence, we can ascertain from figure 6 that the point at which the
434 upper bound crosses the zero difference line to become negative, is the approximate number of
435 SSWs that are required to obtain the required level of statistical significance (see the respective
436 coloured vertical lines).

437

438 In terms of the NAM index, it can be seen that at the 90%, 95% and 99% levels, the number of
439 DW SSWs required is ~ 55 , 75 and 115, respectively (in addition to the same number of NDW
440 SSWs). For $F^{(z)}$, the numbers required are slightly less (~ 40 , 50, and 85), and for Z over the

441 Siberian high sector, the numbers are slightly less again (~ 35 , 45, and 70). This suggests that
442 the tropospheric precursor which most efficiently discriminates DW from NDW events is the
443 strength of the 700-hPa height anomaly over the Siberian High sector. In all three cases, even at
444 the 90% level, the number of DW and NDW SSWs required separately to find such precursor
445 anomalies, is more than double that of the observed number of SSWs in even the JRA-55 re-
446 analysis (which has one of the largest numbers of SSWs among contemporary reanalysis datasets).

447

448 *b. Robustness of these Precursors*

449 The previous section identified tropospheric precursors that appear to distinguish DW and
450 NDW SSWs. We test the robustness of the zonal-mean NAM precursors by comparing the NAM
451 shown in figure 1 with that of randomly-selected tropospheric events which are independent of a
452 SSW (figure 7). The latter allows us to test whether the precursor anomalies to SSWs we have
453 found are simply related to random tropospheric variability. Additionally, we have also tested
454 the robustness to different DW definitions but direct the reader to the supplementary material
455 for figures and analysis. In order to calculate this random composite, we removed each SSW
456 event and its surrounding 100 days (hence, 101 days total for each event) from the timeseries for
457 each experiment, and then randomly selected a new event, which by construction, is unrelated
458 to a SSW. We define each event as having a negative (Tneg) or positive (Tpos) tropospheric
459 NAM after the 'onset date' by averaging the tropospheric NAM at 500 hPa over lags +10 to +50,
460 yielding 411 Tneg and 551 Tpos events (this is similar to the DW definition of Jucker (2016);
461 see supplementary information). By construction, we are sampling only tropospheric internal
462 variability.

463

464 Whilst the negative NAM signal in the Tneg composite for positive lags arises by construction
465 (a), the NAM is also negative at negative lags, due to the persistence of the NAM index. The
466 opposite is evident in the Tpos composite (b), although with a larger amplitude. This is due to the
467 fact that the tropospheric NAM index is on average slightly positive when all SSWs are removed.
468 This yields Tneg-Tpos differences which are significantly negative at all lags (c), and which are
469 qualitatively similar to that found in the DW-NDW differences (but with differing magnitudes;
470 compare with figure 1d). However, we note that these events are randomly chosen and the onset
471 date has no influence on the tropospheric NAM; indeed, the onset date could be randomly chosen
472 to either occur at the start, in the middle, or at the end of the lifecycle of the negative tropospheric
473 NAM event, which when averaged over all 962 events, would conceivably give a composite
474 similar to that shown in figure 7. In fact, upon reselecting events hundreds of times, similar
475 composites are found. Nevertheless, this viscerally highlights that the differences at positive lags
476 in the troposphere are entirely there by construction.

477

478 We now examine the latitude-longitude differences between Tneg and Tpos for the random
479 tropospheric events. Figure 8 shows the GPH anomalies at 700 hPa for the DW and NDW SSW
480 events (left column; reproduced from figure 3a,b), the Tneg and Tpos events (middle column),
481 and the differences DW-Tneg (right column, top) and NDW-Tpos (right column, bottom). The
482 Tneg events show overall much weaker anomalies than the DW SSW events with negative
483 anomalies at midlatitudes associated with a localised trough over the North Pacific basin and a
484 smaller-valued trough over the North Atlantic basin, and positive anomalies further poleward.
485 This yields DW-Tneg differences with a high slightly northwest of the climatological Siberian
486 High and a low slightly to the northeast of the climatological Aleutian Low, similar to figure 3c
487 due to the dominance of the SSW composites. In terms of the Tpos events, there is also a more

488 annular structure, but of opposite sign to the Tneg events, yielding annular and opposite-signed
489 differences to DW-Tneg. The differences between the randomly-selected events and the precursor
490 anomalies present in the DW and NDW SSWs at negative lags allows us to conclude that the
491 enhanced wave forcing we have found at the lower levels is a robust feature and not present due
492 to random tropospheric variability.

493

494 *c. Relationship between SSW Frequency and Precursory Extreme Wave-Activity*

495 Section 3a and 3b have demonstrated that in a large composite of SSWs, tropospheric features
496 before the SSW clearly differentiate between SSWs which have a DW impact and those which do
497 not. However, in order to not overstate the importance of tropospheric precursory features evident
498 in such composites, we now examine the spread of individual SSWs and see how many events,
499 both DW and NDW, show evidence of such precursors.

500

501 Figure 9a-c shows scatter graphs of $F^{(z)}$ (filtered for planetary-wave 1, averaged over 45-75°N
502 and standardised as in figure 2) at three different levels averaged over lags -15 to -1, against the
503 NAM index at 10 hPa averaged over lags +1 to +10. We note that the patterns are not sensitive to
504 slight changes in the earlier lag for $F^{(z)}$. $F^{(z)}$ is filtered for wave-1 as this wavenumber appears
505 to play the largest role in the composites shown in figure 2. We note that the window for $F^{(z)}$
506 used here is shorter than that used in Polvani and Waugh (2004) who found that a time-integrated
507 upward flux over 40 days at 150 hPa gave the best correlation. At all three levels (100, 300, and
508 700 hPa), the correlation coefficients are negative indicating that enhanced wave activity gives
509 rise to a weaker Polar Vortex. However, the overall correlation coefficients are maximised at 100
510 hPa (-0.54), become weaker at 300 hPa (-0.46) and reduce substantially at 700 hPa (-0.33). At all

511 three levels, the correlation coefficients are statistically significant ($p \ll 0.01$), which, given the
512 relatively small correlation coefficient at 700 hPa, is likely due to the large sample size. Upon
513 splitting into DW and NDW events, and calculating the lines of best fit for each, it can be seen
514 that the respective correlation coefficients are also both very similar at 100 hPa (-0.50 and -0.56).
515 300 hPa (-0.43 and -0.47) and at 700 hPa (-0.28 and -0.34). The scatter about the lines of best
516 fit, particularly at the lower two levels, is indicative of the high degree of variability in the winter
517 troposphere and stratosphere. The composite mean for both event types (large squares) indicate
518 that for DW events, there is a slightly larger upward wave-activity flux at all levels preceding the
519 SSW, which results in a more negative 10-hPa NAM.

520

521 The decline in the correlation between the stratospheric NAM and the vertical component
522 of the EP flux as one analyses the EP flux closer to the surface is consistent with the recent
523 papers by Birner and Albers (2017) and also de la Camara et al. (2017). Specifically, Birner and
524 Albers (2017) found that 25% of SSWs in the relatively short reanalysis record were preceded by
525 extreme lower-tropospheric wave events (LTWEs; 700 hPa). We here further update this statistic
526 using our large ensemble of SSWs. We define a SSW to be preceded by extreme wave activity
527 at a given level if the deseasonalised 11-day running-mean averaged $F^{(z)}$ exceeds the 2-standard
528 deviation threshold at least once in the preceding 10 days (this 10-day window was found to be
529 appropriate by Sjoberg and Birner 2012; Birner and Albers 2017). This is performed separately
530 for waves 1 and 2, and in order to avoid double counting, if a given SSW event is preceded by
531 both extreme wave-1 and wave-2 fluxes, the wavenumber with the largest $F^{(z)}$ value is used to
532 define the dominant wavenumber preceding the SSW.

533

534 Hence we plot in figure 10a, the percentage of SSWs which are preceded by extreme upward
535 wave activity as a function of height for wave 1 (green), wave 2 (red) and wave 1 and wave
536 2 together (blue). The overall profile for wave 1 shows that 45% of SSWs are preceded by at
537 least one day of extreme wave-1 activity at 100 hPa. This figure decreases fairly rapidly with
538 decreasing height with 23% of SSWs being preceded by extreme wave-1 activity at 700 hPa. For
539 wave-2 on the other hand, the percentage of SSWs which are preceded by extreme wave activity
540 at 100 (700) hPa is much smaller than wave-1 with values of 14% (8%). Perhaps most tellingly, if
541 we combine the two then 31% of SSWs are preceded by extreme wave activity at 700 hPa which
542 is similar to the 25% observed by Birner and Albers (2017) using ERA Interim reanalysis. At 100
543 hPa, this combined percentage rises to $\sim 60\%$.

544

545 While this result indicates that roughly one third of SSWs are preceded by extreme wave activity
546 in the lower troposphere, additional insight as to the usefulness of tropospheric wave activity for
547 predicting a SSW can be obtained by examining the number of lower-tropospheric wave events
548 (LTWEs) which are followed by SSWs. We define such a LTWE if the 11-day running-mean
549 averaged $F^{(z)}$ at 700 hPa exceeds the 2-standard deviation threshold during wintertime (Oct-April).
550 The difference in the number of days between two consecutive LTWEs must be greater than or
551 equal to 10 days. If there is any overlap between any wave-1 and wave-2 events within 10 days,
552 then as before, the larger-valued wavenumber is assumed to be dominant. This yields 1374 (1311)
553 extreme wave-1 (wave-2) LTWEs¹. The percentage of LTWEs which are followed by a SSW

¹We note that this definition is slightly different to that used in Birner and Albers (2017) who define a start and end date for a LTWE as the first exceedance of 2 standard deviations and the subsequent first drop below 2 standard deviations, respectively. Then, no other LTWE can be defined in the 20 days following the end date (personal communication). A SSW is determined to follow the LTWE if it occurs within 10 days of the end date. Nevertheless, our results are insensitive to this definition, as in our analysis this definition yields 2626 (1338 wave-1 and 1288 wave-2) independent LTWEs, with 27% of SSWs being preceded by a LTWE in this way (compare with 31%).

554 is then calculated from the SSWs shown above and the number of LTWEs. The corresponding
555 percentages are inset into the panels in figure 10a; 16% (6%) of 700-hPa wave-1 (wave-2) LTWEs
556 are followed by a SSW, which together indicates that 11% of LTWEs appear to be followed by a
557 SSW event.

558

559 In figure 10b, the percentage of SSWs which are preceded by extreme wave activity at each level
560 and which subsequently go on to be either DW or NDW propagating is shown. By construction,
561 the DW and NDW profiles when summed at each level, equal 100%. The DW profile maximises
562 in the lower troposphere (below ~ 400 hPa) suggesting that the presence of extreme wave activity
563 in the lower troposphere appears to be a better indicator of whether the SSW will go on to be
564 DW propagating than such extreme wave activity at higher levels. Indeed, the percentage of
565 SSWs which are preceded by extreme wave activity at 700 hPa and which are subsequently
566 DW propagating is 64% (and conversely 36% for NDW propagation). Hence, in a probabilistic
567 sense, there is a 28% difference between DW- and NDW-propagating SSWs and the tropospheric
568 wave activity which occurs prior to it (consistent with section 3a). However, given that a high
569 percentage of SSWs which are preceded by extreme lower-tropospheric wave activity are NDW
570 propagating, one would not be able to make a deterministic prediction at the onset of whether a
571 given SSW will be DW or NDW propagating.

572

573 We note that the same analysis was also performed using the standardised anomalies over the
574 Siberian High sector ($50-80^{\circ}\text{N}$, $60-90^{\circ}\text{E}$) at 700 hPa. The percentages were around half of those
575 shown in figure 10, with 16% of the total number of SSWs being preceded by such extreme (>2
576 standard deviations) anomalies. The percentage of SSWs preceded by such anomalies which
577 then go on to be DW (NDW) propagating is 62% (38%). Hence despite figure 6 indicating

578 that examining the GPH anomalies over the Siberian High sector may be a more robust way to
579 examine the DW influence of SSWs, these percentages indicate that instead $F^{(z)}$ may be a better
580 indicator.

581

582 *d. Precursors to Splits and Displacements*

583 So far we have only focussed on the precursors to SSWs identified using the Charlton and
584 Polvani (2007) approach. Here we examine the precursors associated with splits and displace-
585 ments identified using the method of Seviour et al. (2013). Additionally, in light of recent studies
586 which have found differing results with regards to which type of event has the most noticeable
587 surface impact after the onset date (Mitchell et al. 2013; Seviour et al. 2013; Maycock and
588 Hitchcock 2015), we again use the DW definition of Karpechko et al. (2017) to examine the DW
589 influence of both splits and displacements.

590

591 Figure 11 shows the height-time evolution of the NAM index divided into displacements
592 (left column) and splits (middle column) and subdivided further into the total (top row), DW-
593 propagating (middle row) and NDW-propagating (bottom row). Also shown are the differences
594 (right column) for displacements-splits (top), DW-NDW displacements (middle) and DW-NDW
595 splits (bottom). In the total composites, clear significant differences between displacements
596 and splits can be seen in both the stratosphere and in the troposphere. In the stratosphere, the
597 displacements are stronger than the splits, up until lag +50. In particular, in the middle-to-upper
598 stratosphere the displacements are nearly twice as strong. In the troposphere, whilst the dis-
599 placement events have a stronger long-term influence up until lag +45, the splits have a more
600 barotropic nature at the onset with an instantaneous response near the surface, which dissipates

601 after \sim lag +5. The barotropic nature at the onset is in agreement with the more likely role of the
602 barotropic mode for split SSWs (Esler and Scott 2005). Prior to the onset date, the splits show
603 clear tropospheric negative anomalies extending back to lag -45 which are stronger than for the
604 displacements.

605

606 Upon subdividing into DW (middle row) and NDW (bottom row) events, the splits and
607 displacements broadly show similar results to those found using the wind reversal criterion
608 (figure 1) with slightly stronger negative NAM anomalies in the middle to upper stratosphere as
609 well as longer-persisting anomalies in the lower stratosphere for DW events. This yields therefore,
610 similar DW-NDW composite differences at positive lags to figure 1. However, at negative lags,
611 the splits have much stronger negative tropospheric and lower-stratospheric precursors than
612 the displacements, extending back to lag -55 and becoming stronger around lag -25 for the
613 DW events, but weaker anomalies extending back to lag -30 for the NDW splits. The DW
614 displacements on the other hand show very similar anomalies to the total (a), and the NDW
615 displacements show evidence of positive tropospheric anomalies up to two weeks before the
616 onset (and weakly negative anomalies before that). Overall, this gives similar-valued DW-NDW
617 differences at negative lags, except that the splits have negative differences which extend further
618 back to lag -30 and also extend into the stratosphere.

619

620 As before, we now examine the regional differences in order to understand these tropospheric
621 precursors. Figure 12 shows the same as the PC anomalies in figure 3 except for Z at 700 hPa
622 for the (top) displacement and (bottom) split events. Note that we don't show the ONS and REC
623 stage in this plot as they are similar to those in figure 3. For the displacement events, there are
624 negative anomalies over the Northwestern Pacific and positive anomalies over Northern Europe

625 and Siberia. These two anomalous centres project onto the climatological wave-1 centres of
626 action (green contours), and in particular, the positive anomaly over Northern Europe/Siberia is
627 more positive for the DW events, indicating similarly to figure 3, an increase in upward wave-1.
628 Also over the subtropical North Pacific, there is a band of positive anomalies projecting onto the
629 eastern flank of the climatological wave-1 Aleutian Low. These anomalies are more positive under
630 NDW events and hence yield negative differences over the Aleutian Low sector. This subtropical
631 band of positive anomalies in conjunction with the negative anomalies further poleward, yield
632 a dipole over the Pacific basin leading to possible meridional shifts in the East Pacific Jet (e.g.,
633 Nishii et al. 2010; Dai and Tan 2016; Bao et al. 2017).

634
635 For the split events (bottom), the anomalies at this level show more of a wave-2 structure,
636 with an intensification of the highs and lows of the climatological wave-2 (green contours).
637 In particular, there are negative anomalies over the North Pacific, over the North Atlantic and
638 Western Europe, along with positive anomalies over Siberia and Eastern Europe. In general,
639 these anomalies are stronger for the DW events, as indicated by the difference composite. The
640 differences also show evidence of an intensification of the climatological wave-1.

641
642 We now plot the height-time evolution of $F^{(z)}$ for displacement events (figure 13) and split
643 events (figure 14) in order to determine the vertical extent of the wave-1 (top row) and wave-2
644 (bottom row) anomalies from the troposphere into the stratosphere. As in figure 2, the anomalies
645 are standardised by their standard deviation at each pressure level. For the displacements, the
646 wave-1 anomalies are generally similar to those in the wave 1-2 composite shown in figure 2.
647 For DW events, there is enhanced upward wave-1 compared to NDW events, which propagates
648 up from 700 hPa into the stratosphere peaking close to the onset date. After the onset, the wave

649 activity is generally suppressed as shown by negative anomalies in both the DW and NDW events,
650 although positive (upward) anomalies do persist in the upper troposphere to lower stratosphere for
651 ~ 5 -10 days after the onset. The negative anomalies for the NDW events are of significantly larger
652 magnitude. Note that the other wavenumbers contribute negligibly to the $F^{(z)}$ flux and hence we
653 do not include them here, for brevity.

654

655 For split events (figure 14), we can see that they are generally preceded by upward wave-1
656 and wave-2 anomalies which propagate up from 700 hPa and peak in the stratosphere. As in the
657 displacements, the standardised anomalies are larger in the stratosphere than in the troposphere.
658 This is the case for both DW and NDW events, although there is actually slightly less upward
659 wave-2 at the onset for the DW events (f; opposite to Nakagawa and Yamazaki 2006). However,
660 those which propagate DW to the troposphere are on average preceded by enhanced anomalous
661 upward wave-1 into the stratosphere (c). In the wave-1 difference (c) it can be seen that this
662 enhanced upward wave-1 for DW events starts around lag -20 and persists through the onset date
663 until around lag +10. Even though split events are generally associated with wave-2 anomalies
664 in the upward flux (as shown in d,e), this result indicates that wave-1 may also play a role in the
665 DW influence. Similar to the displacements, there are enhanced upward tropospheric wave-2
666 anomalies for the DW events after the onset date.

667

668 **4. Summary and Discussion**

669 Using a series of 40 integrations of the GEOSCCM model, we have (1) identified and anal-
670 ysed the frequency of tropospheric precursory features to SSWs (generally, and for splits and
671 displacements) which appear to manifest as zonally-varying wave patterns that project onto the

672 climatological stationary planetary centres, extending the recent observational study of Birner and
673 Albers (2017), and (2) examined the differences in such precursors between so-called downward
674 (DW) and nondownward (NDW) propagating SSWs. To do this we identified a large compendium
675 of SSWs across all 40 runs using the definition of Charlton and Polvani (2007). This yielded a
676 ratio of approximately 0.61 SSWs per year (~ 950 in ~ 1600 years) which were then classified as
677 DW and NDW-propagating using a variety of recently-developed DW definitions (Jucker 2016;
678 Runde et al. 2016; Karpechko et al. 2017).

679

680 For the SSWs in general, there is an enhanced upward flux of wave activity into the stratosphere
681 from the troposphere preceding the SSW onset. In a composite sense, the enhanced wave activity
682 appears to originate in the lower troposphere (figures 2-5 and 13-14), although relative to its local
683 standard deviation, the anomalies in the stratosphere are at least twice as large as those in the
684 troposphere, in agreement with similar composites in Jucker (2016) and Birner and Albers (2017).
685 This occurs as a projection of the anomalies onto the climatological centres of action, associated
686 with a deepening of the Aleutian Low and a strengthening of the Siberian High and yielding
687 an enhanced upward wave-1 flux. The enhancement of upward wave-1 activity prior to the
688 onset, followed by the subsequent reduction at later times is in agreement with the observational
689 composites of Limpasuvan et al. (2004) using reanalysis data.

690

691 Recent studies by Jucker (2016), Birner and Albers (2017) and de la Camara et al. (2017)
692 found that anomalous upward fluxes of lower-tropospheric wave activity were not a necessary or
693 sufficient precursor to SSW events, given that only one quarter of SSWs in the period covered
694 by ERA-Interim were preceded by such wave events. Instead, they found that the state of the
695 stratosphere prior to the onset date played a much more important role in determining the occur-

696 rence of a SSW. The stratospheric state may be in a preferable configuration to take advantage of
697 the climatologically-large tropospheric reservoir of wave activity and encourage an anomalous
698 upward wave flux across the tropopause. Our results in section 3c agree well with the results of
699 Birner and Albers (2017), despite the shortness of the observational record, as 31% of SSWs are
700 here found to be preceded by extreme lower-tropospheric (700-hPa) wave activity (figure 10).

701

702 The number of SSWs which were preceded by extreme wave activity increases rapidly up to
703 100 hPa (~60%). Given that at high latitudes the 100-hPa surface is already well within the
704 vortex (de la Camara et al. 2017), this is perhaps expected. Furthermore, the correlations between
705 the vertical wave flux (which is again maximised at 100 hPa) and the strength of the Polar Vortex
706 at 10 hPa, reduce substantially closer to the surface (figure 9). This is indicative of the fact that
707 even in the presence of lower-tropospheric wave activity, the high degree of internal atmospheric
708 variability can easily prevent such wave activity from propagating upward into the stratosphere.
709 Indeed, it still remains to be seen how even in the presence of extreme tropospheric wave fluxes,
710 the stratosphere can (or cannot) take advantage of such anomalous wave fluxes. However, our
711 study cannot shed light on the ingredient which allows for this.

712

713 In the case of DW-propagating SSWs, we find evidence of both significantly enhanced
714 zonal-mean and regional tropospheric precursors, compared to the NDW SSWs in the composites
715 shown in figures 1- 5. In terms of the zonal-mean, negative NAM anomalies were found to
716 exist throughout the troposphere prior to the onset date for DW events, with negative DW-NDW
717 differences extending as far back as lag -40 (see figure 1). NAM precursors were also found
718 previously using large numbers of simulated SSW events (e.g., Jucker 2016; Karpechko et al.
719 2017). However, as aforementioned, such NAM precursors have been shown to be model-,

720 and configuration dependent (Gerber et al. 2010). This is consistent with Black and McDaniel
721 (2004) who observed that the determination of the DW propagation of a SSW depended on the
722 pre-existing tropospheric state, with a pre-existing positive NAM-like state being associated with
723 NDW SSWs, and vice versa. Note that using three of the four recently-proposed DW definitions
724 (Runde et al. 2016; Jucker 2016; Karpechko et al. 2017), yields similar precursory features (see
725 supplementary information for details and a discussion of the fourth definition which yields
726 different results).

727

728 Further, enhanced upward zonal-mean wave-activity fluxes ($F^{(z)}$) were also found (figure 2) to
729 precede DW SSWs extending back to around lag -25. These standardised anomalies spanned the
730 depth of the troposphere and intensified in the stratosphere above 200 hPa. By splitting the SSWs
731 according to the magnitude of the $F^{(z)}$ anomalies prior to the onset date rather than according to
732 the magnitude of the NAM after the onset, it was found that on average, those events with larger
733 $F^{(z)}$ led to a more negative tropospheric NAM signal after the onset (figure 4).

734

735 In a regional sense, there appear to be differences between DW and NDW-propagating SSWs
736 in the geopotential height in the troposphere and lower stratosphere (figures 3-5 and 8), which
737 strengthen the wave anomalies already associated with the onset of the SSW. The regional
738 differences are particularly large over Northern Europe and Siberia, with a strengthening of
739 the climatological Siberian High under DW events. We note that such anomalies over the
740 Siberian-High sector prior to DW-propagating SSW events were also found in observations by
741 Nakagawa and Yamazaki (2006) using the 45-year ERA-40 reanalysis dataset.

742

743 Previous work has showed a wide disparity in the sign of the tropospheric NAM signal before
744 SSWs (see figure 10 of Gerber et al. 2010). With the availability of 900+ SSWs, we more clearly
745 see this negative NAM precursor, although at least 55 DW and 55 NDW events are needed
746 before this NAM feature becomes robust (figure 6; although note that only 35-40 DW and NDW
747 events separately, are required to find robust differences in $F^{(z)}$ and Z). Indeed, in only a handful
748 of the individual members of the 40-member ensemble are such tropospheric NAM precursors
749 present (not shown), suggesting that the diversity evident in Gerber et al. (2010) arises not only
750 from peculiarities of the various models but also from internal variability. Note that this is also
751 in agreement with the work of Gerber et al. (2009) and Hitchcock and Simpson (2014) who
752 suggested that the tropospheric response to a SSW consisted of a forced tropospheric component
753 (by the SSW) and a stochastic component which is independent of the SSW above. Indeed, in their
754 runs, they found that a given SSW event may or may not influence the troposphere, depending
755 on tropospheric natural variability which can act to mask any actual DW stratospheric signal. As
756 our analysis indicates that at least 55 SSWs of each type are required before the NAM-precursor
757 effect becomes salient, it shows that internal tropospheric variability can indeed mask any forced
758 signal from the stratosphere. Nevertheless, our results also indicate that the forced signal from
759 the stratosphere is stronger on average if the precursory wave flux from the troposphere is stronger.

760
761 Examining the numbers of SSWs which are preceded by extreme lower-tropospheric wave
762 activity and go on to be DW or NDW propagating gives an idea as to how useful such precursory
763 wave activity may be in predicting the tropospheric impact following a SSW. Indeed, of the 296
764 SSWs which were preceded by such wave activity, 64% (36%) subsequently went on to be DW
765 (NDW) propagating. This enhances the probabilistic prediction of tropospheric impacts following
766 a SSW as it suggests that if a given SSW was preceded by extreme lower-tropospheric wave

767 activity, then one could say at the onset, that there is a greater likelihood that it will propagate
768 DW to the troposphere. However, given that a relatively high percentage of SSWs were also
769 preceded by such wave activity and went on to be NDW propagating, one would not be able to
770 make a deterministic prediction before the onset of whether a given SSW will be DW or NDW
771 propagating. Nevertheless, these percentages augment themselves with similar percentages shown
772 in Karpechko et al. (2017, see their figure 5) whose results suggested that the likelihood of a SSW
773 having a DW tropospheric impact depends on the sign and magnitude of the lower-stratospheric
774 NAM index and $F^{(z)}$ just after the onset date; in particular, the more negative the 150-hPa NAM is
775 at lags 0-4 following the SSW, the more likely it is to propagate DW at later lags.

776
777 We also compared the results to those obtained using composites of randomly-selected tro-
778 pospheric events, which by construction, were chosen to be unrelated to the SSW above (see
779 section 3b). In a zonal-mean, the composites for the DW and NDW SSWs and for the negative
780 (T_{neg}) and positive (T_{pos}) random tropospheric events were remarkably similar at all lags
781 (figure 7), albeit with changes in magnitude. The replicability of the tropospheric zonal-mean
782 NAM at both positive and negative lags using random events based solely on the behaviour of the
783 troposphere, suggests exhibiting caution to just using the NAM to examine the DW influence of a
784 SSW event, as it can conceal much of the regional information that is important for understanding
785 the precursors.

786
787 However, the regional precursors, which were found to be associated with upward planetary
788 wave-1 forcing for the SSW events, were very different for the random composites, instead having
789 a weak, annular structure (figure 8). Because of the differences in the regional tropospheric
790 precursory features between SSW events and randomly-selected events, we conclude that the

791 precursors here found are robust and that there is a difference prior to DW and NDW SSWs other
792 than just random tropospheric variability.

793

794 The converse to examining the proportion of SSWs (either DW or NDW propagating) which are
795 preceded by extreme lower-tropospheric wave activity is to consider the proportion of such events
796 which are followed by a SSW within 10 days. In total, 11% of the identified lower-tropospheric
797 wave events (16% of wave-1 and 6% of wave-2) were followed by a SSW. Despite this figure
798 being twice as large as the observed 6% of tropospheric blocks which are followed by a SSW
799 event in 44 years of reanalysis data (Martius et al. 2009), we stress that it is impractical to forecast
800 SSWs based solely on identifying extreme tropospheric wave events (e.g., Birner and Albers
801 2017).

802

803 We finally examined the evolution of the troposphere and stratosphere associated with split
804 and displacement SSW events. We found that: 1) displacements tend to have a longer-term
805 tropospheric influence, and 2) splits have a more barotropic influence at the onset date (figure 11).
806 The former is in agreement with Maycock and Hitchcock (2015) using a large sample of SSWs
807 from a long model integration and the method of Seviour et al. (2013) to classify events. However,
808 their results were not robust as using a different classification method, yielded different results.
809 Regarding split SSWs, the barotropic influence is in agreement with the barotropic mode leading
810 to a split SSW (Esler and Scott 2005; Matthewman et al. 2009; Seviour et al. 2016). However,
811 these results overall disagree with studies by Mitchell et al. (2013), Seviour et al. (2013),
812 O'Callaghan et al. (2014) and Lehtonen and Karpechko (2016) who found that splits have a
813 larger tropospheric influence than displacements in reanalysis data lasting up until lag +60. The
814 disagreement may be related to the differences in sample sizes which is an order of magnitude

815 larger in our study. Indeed, we created composites for each individual experiment (not shown),
816 and in a handful of the 40 ensemble members, composites are qualitatively similar to Mitchell
817 et al. (2013). However, we note that our results are more in agreement with Seviour et al. (2016),
818 who used 13 stratosphere-resolving models from the fifth Coupled Model Intercomparison Project
819 (CMIP5) ensemble and found that despite splits exhibiting a slightly stronger signal over the
820 North Atlantic for up to one month after the SSW, the largest and most significant differences
821 were associated with displacements over Siberia. We note that our results therefore, are also
822 slightly in disagreement with Karpechko et al. (2017), who in their large ensemble of SSWs
823 obtained from a chemistry-climate model, instead found indistinguishable differences between
824 the two types of events.

825

826 We also found that in general, the splits and displacements were associated with enhanced
827 upward wave-2 and wave-1 forcing respectively (figures 13-14; e.g., Andrews et al. 1987;
828 Nakagawa and Yamazaki 2006; Liu et al. 2014; Lehtonen and Karpechko 2016) extending into
829 the middle-to-lower troposphere, although we note that there was enhanced wave-1 present for
830 both types. Further, those splits and displacements which propagate DW to the troposphere were
831 associated with even further enhanced wave-1 fluxes at negative lags as compared to NDW-
832 propagating events. The enhanced wave-2 forcing for the splits was more barotropic, occurring
833 closer to the onset date, than for the enhanced wave-1 forcing. The near-barotropic wave-2 nature
834 closer to the onset in association with the larger percentage of SSWs being preceded by extreme
835 lower-tropospheric wave-1 rather than wave-2 fluxes (figure 10) suggest that split SSWs may be
836 more nonlinear and thus potentially more difficult to predict.

837

838 The results in this paper indicate that the strength of the wave forcing both prior to and during
839 the SSW onset and the subsequent strength of the SSW, may play a role in the DW influence of
840 the SSW. However, as mentioned previously, the results only show evidence of an enhancement
841 in probabilistic forecasts of the DW influence; deterministically one could not say if a given
842 SSW event will have such an influence. Hence, given the statistical nature of our analysis, we
843 cannot establish whether the precursor patterns associated with DW-propagating SSWs identified
844 here, play a causal role in the tropospheric impact. As this paper only focusses on the output
845 from a single model, future work using observations and/or integrations using different models is
846 required to determine whether the enhanced wave-1 activity, and zonal structure of the precursors
847 (e.g., the enhanced Siberian High), play a role in the mechanism, and if so, how.

848
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858 **References**

859 Andrews, D. G., J. R. Holton, and C. B. Leovy, 1987: *Middle Atmosphere Dynamics*. Academic
860 Press, 489 pp.

- 861 Andrews, D. G., and M. E. McIntyre, 1978: Generalized Eliassen-Palm and Charney-Drazin the-
862 orem for waves on axisymmetric mean flows in compressible atmospheres. *J. Atmos. Sci.*, **35**,
863 175–185.
- 864 Aquila, V., W. H. Swartz, D. W. Waugh, P. R. Colarco, S. Pawson, L. M. Polvani, R. S. Stolarski,
865 and D. W. Waugh, 2016: Isolating the roles of different forcing agents in global stratospheric
866 temperature changes using model integrations with incrementally added single forcings. *J. Geo-
867 phys. Res. Atmos.*, **121**, 8067–8082.
- 868 Baldwin, M. P., and T. J. Dunkerton, 2001: Stratospheric harbingers of anomalous weather
869 regimes. *Science*, **294**, 581–584.
- 870 Baldwin, M. P., and D. W. J. Thompson, 2009: A critical comparison of stratosphere-troposphere
871 coupling indices. *Q.J.R. Meteorol. Soc.*, **135**, 1661–1672.
- 872 Bao, M., D. L. Hartmann, and P. Ceppi, 2017: Classifying the tropospheric precursor patterns of
873 sudden stratospheric warmings. *Geophys. Res. Lett.*, **44**.
- 874 Birner, T., and J. R. Albers, 2017: Sudden stratospheric warmings and anomalous upward wave
875 activity flux. *Sci. Onl. Lett. Atmos.*, **13A**, 8–12.
- 876 Black, R. X., and B. A. McDaniel, 2004: Diagnostic case studies of the northern annular mode. *J.*
877 *Clim.*, **17**, 3990–4004.
- 878 Butler, A. H., D. J. Seidel, S. C. Hardiman, N. Butchart, T. Birner, and A. Match, 2015: Defining
879 sudden stratospheric warmings. *Bull. Amer. Meteor. Soc.*, **96**, 1913–1928.
- 880 Charlton, A. J., and L. M. Polvani, 2007: A new look at stratospheric sudden warmings. Part I:
881 Climatology and modeling benchmarks. *J. Clim.*, **20**, 449–469.

882 Charney, J. G., and P. G. Drazin, 1961: Propagation of planetary scale disturbances from the lower
883 into the upper atmosphere. *J. Geophys. Res.*, **66** (1), 83–109.

884 Cohen, J., and J. Jones, 2011: Tropospheric precursors and stratospheric warmings. *J. Clim.*, **24**,
885 6562–6572.

886 Dai, Y., and B. Tan, 2016: The western pacific pattern precursor of major stratospheric sud-
887 den warmings and the ENSO modulation. *Env. Res. Lett.*, **11** (12), URL [http://stacks.iop.org/
888 1748-9326/11/i=12/a=124032](http://stacks.iop.org/1748-9326/11/i=12/a=124032).

889 de la Camara, A., J. A. Albers, T. Birner, R. R. Garcia, P. H. Hitchcock, D. E. Kinnison, and A. K.
890 Smith, 2017: Sensitivity of sudden stratospheric warmings to previous stratospheric conditions.
891 *J. Atmos. Sci.*, **74**, 28572877.

892 Domeisen, D. I. V., L. Sun, and G. Chen, 2013: The role of synoptic eddies in the tropospheric
893 response to stratospheric variability. *Geophys. Res. Lett.*, **40**, 4933–4937.

894 Esler, J. G., and R. K. Scott, 2005: Excitation of transient rossby waves on the stratospheric polar
895 vortex and the barotropic sudden warming. *J. Atmos. Sci.*, **62**, 3661–3682.

896 Garfinkel, C. I., A. Gordon, L. Oman, F. Li, S. Davis, and S. Pawson, 2017: Nonlinear response of
897 tropical lower stratospheric temperature and water vapor to ENSO. *Atmos. Chem. Phys. Disc.*,
898 **18**, 4597–4615.

899 Garfinkel, C. I., D. L. Hartmann, and F. Sassi, 2010: Tropospheric precursors of anomalous north-
900 ern hemisphere stratospheric polar vortices. *J. Clim.*, **23**, 3282–3299.

901 Garfinkel, C. I., D. W. Waugh, and L. M. Polvani, 2015: Recent hadley cell expansion: The role
902 of internal atmospheric variability in reconciling modeled and observed trends. *Geophys. Res.
903 Lett.*, **42**, 10 824–10 831.

- 904 Gerber, E. P., C. Orbe, and L. P. Polvani, 2009: Stratospheric influence on the tropospheric circu-
905 lation revealed by idealized ensemble forecasts. *Geophys. Res. Lett.*, **36**.
- 906 Gerber, E. P., and Coauthors, 2010: Stratosphere-troposphere coupling and annular mode variabil-
907 ity in chemistry-climate models. *J. Geophys. Res. Atmos.*, **115**.
- 908 Hitchcock, P., and P. H. Haynes, 2016: Stratospheric control of planetary waves. *Geophys. Res.*
909 *Lett.*, **43**, 11 884–11 892.
- 910 Hitchcock, P., T. G. Shepherd, and G. L. Manney, 2013: Statistical characterization of arctic polar-
911 night jet oscillation events. *J. Atmos. Sci.*, **26**, 20962116.
- 912 Hitchcock, P., and I. R. Simpson, 2014: The downward influence of stratospheric sudden warm-
913 ings. *J. Atmos. Sci.*, **71**, 3856–3876.
- 914 Jucker, M., 2016: Are sudden stratospheric warmings generic? insights from an idealized gcm. *J.*
915 *Atmos. Sci.*, **73**, 5061–5080.
- 916 Karpechko, A. Y., P. Hitchcock, D. H. W. Peters, and A. Schneidereit, 2017: Predictability of
917 downward propagation of major sudden stratospheric warmings. *Q. J. R. Meteorol. Soc.*, **143**,
918 1459–1470.
- 919 Kidston, J., A. A. Scaife, S. C. Hardiman, D. M. Mitchell, N. Butchart, M. P. Baldwin, and
920 L. J. Gray, 2015: Stratospheric influence on tropospheric jet streams, storm tracks and surface
921 weather. *Nat. Geos.*, **8**, 433–440.
- 922 Kodera, K., H. Mukougawa, P. Maury, M. Ueda, and C. Claud, 2016: Absorbing and reflect-
923 ing sudden stratospheric warming events and their relationship with tropospheric circulation. *J.*
924 *Geophys. Res. Atmos.*, **121**, 80–94.

925 Lehtonen, I., and A. Y. Karpechko, 2016: Observed and modeled tropospheric cold anomalies
926 associated with sudden stratospheric warmings. *J. Geophys. Res. Atmos.*, **121**, 1591–1610.

927 Limpasuvan, V., D. W. Thompson, and D. L. Hartmann, 2004: The life cycle of the northern
928 hemisphere sudden stratospheric warmings. *J. Clim.*, **17**, 2584–2596.

929 Liu, C., K.-F. Tian, G. L. Manney, N. J. Livesey, Y. L. Yung, and D. E. Waliser, 2014: Northern
930 hemisphere mid-winter vortex-displacement and vortex-split stratospheric sudden warmings:
931 Influence of the Madden-Julian Oscillation and Quasi-Biennial Oscillation. *J. Geophys. Res.*
932 *Atmos.*, **119**, 12,599–12,620.

933 Marshall, A. G., and A. A. Scaife, 2010: Improved predictability of stratospheric sudden warming
934 events in an atmospheric general circulation model with enhanced stratospheric resolution. *J.*
935 *Geophys. Res.*, **115** (D16114).

936 Martineau, P., and S.-W. Son, 2015: Onset of circulation anomalies during stratospheric vortex
937 weakening events: The role of planetary-scale waves. *J. Clim.*, **28**, 7347–7370.

938 Martius, O., L. M. Polvani, and H. C. Davies, 2009: Blocking precursors to stratospheric sudden
939 warming events. *Geophys. Res. Lett.*, **36** (L14806).

940 Matsuno, T., 1971: A dynamical model of the stratospheric sudden warming. *J. Atmos. Sci.*, **28**,
941 1479–1494.

942 Matthewman, N. J., J. G. Esler, A. J. Charlton-Perez, and L. M. Polvani, 2009: A new look at
943 stratospheric sudden warmings. Part III: Polar vortex evolution and vertical structure. *J. Clim.*,
944 **22**, 1566–1585.

945 Maycock, A. C., and P. Hitchcock, 2015: Do split and displacement sudden stratospheric warm-
946 ings have different annular mode signatures? *Geophys. Res. Lett.*, **42**, 10 943–10 951.

947 Mitchell, D., L. J. Gray, J. A. Anstey, M. P. Baldwin, and A. J. Charlton-Perez, 2013: The influence
948 of stratospheric vortex displacements and splits on surface climate. *J. Clim.*, **26**, 2668–2682.

949 Molod, A., M. Takacs, M. Suarez, J. Bacmeister, I.-S. Song, and A. Eichmann, 2012: The geos-5
950 atmospheric general circulation model: Mean climate and development from merra to fortuna:
951 Tech. rep. ser. on global model. and data assimilation. Tech. rep., Nat. Aeronautics and Space
952 Admin., Goddard Space Flight Cent., Greenbelt, Md.

953 Nakagawa, K. I., and K. Yamazaki, 2006: What kind of stratospheric sudden warming propagates
954 to the troposphere? *Geophys. Res. Lett.*, **33 (L04801)**.

955 Nishii, K., H. Nakamura, and Y. Orsolini, 2010: Cooling of the wintertime arctic stratosphere
956 induced by the western pacific teleconnection pattern. *Geophys. Res. Lett.*, **37 (L13805)**.

957 O’Callaghan, A., M. Joshi, D. Stevens, and D. Mitchell, 2014: The effects of different sudden
958 stratospheric warming types on the ocean. *Geophys. Res. Lett.*, **41**, 7739–7745.

959 Pawson, S., R. S. Stolarski, A. R. Douglass, P. A. Newman, J. E. Nielsen, S. M. Frith, and M. L.
960 Gupta, 2008: Earth observing system chemistry-climate model simulations of stratospheric
961 ozone-temperature coupling between 1950 and 2005. *J. Geophys. Res. Atmos*, **113 (D12103)**.

962 Polvani, L. M., and D. W. Waugh, 2004: Upward wave activity flux as a precursor to extreme
963 stratospheric events and subsequent anomalous surface weather regimes. *J. Clim.*, **17 (18)**,
964 3548–3554.

965 Rienecker, M. M., and Coauthors, 2008: The GEOS-5 data assimilation system documentation of
966 versions 5.0.1, 5.1.0, and 5.2.0. *Technical Report Series on Global Modeling and Data Assimi-*
967 *lation*, **27**, 347–367.

- 968 Runde, T., M. Dameris, H. Garny, and D. E. Kinnison, 2016: Classification of stratospheric ex-
969 treme events according to their downward propagation to the troposphere. *Geophys. Res. Lett.*,
970 **43**, 6665–6672.
- 971 Scaife, A. A., and Coauthors, 2012: Climate change projections and stratosphere-troposphere
972 interaction. *Clim. Dyn.*, **38**, 2089–2097.
- 973 Seviour, W. J. M., L. J. Gray, and D. M. Mitchell, 2016: Stratospheric polar vortex splits and
974 displacements in the high-top CMIP5 climate models. *J. Geophys. Res. Atmos.*, **121**, 1400–
975 1413.
- 976 Seviour, W. J. M., D. M. Mitchell, and L. J. Gray, 2013: A practical method to identify displaced
977 and split stratospheric polar vortex events. *Geophys. Res. Lett.*, **40**, 5268–5273.
- 978 Sigmond, M., J. F. Scinocca, V. V. Kharin, and T. G. Shepherd, 2013: Enhanced seasonal forecast
979 skill following stratospheric sudden warmings. *Nat. Geosci.*, **6**, 98–102.
- 980 Sjoberg, J. P., and T. Birner, 2012: Transient tropospheric forcing of sudden stratospheric warm-
981 ings. *J. Atmos. Sci.*, **69**, 3420–3432.
- 982 Smith, D., A. A. Scaife, and B. Kirtman, 2012: What is the current state of scientific knowledge
983 with regard to seasonal and decadal forecasting? *Env. Res. Lett.*, **7** (015602).
- 984 Song, Y., and W. Robinson, 2004: Dynamical mechanisms for stratospheric influences on the
985 troposphere. *J. Atmos. Sci.*, **61**, 1711–1725.
- 986 Thompson, D. W. J., M. P. Baldwin, and J. M. Wallace, 2002: Stratospheric connection to northern
987 hemisphere wintertime weather: Implications for prediction. *J. Climate.*, **15**, 1421–1428.

988 Tomassini, L., E. P. Gerber, M. P. Baldwin, F. Bunzel, and M. J. Giorgetta, 2012: The role of
989 stratosphere-troposphere coupling in the occurrence of extreme winter cold spells over northern
990 europe. *J. Adv. Model. Earth Syst.*, **4**.

991 Tripathi, O. P., and Coauthors, 2014: The predictability of the extratropical stratosphere on
992 monthly time-scales and its impact on the skill of tropospheric forecasts. *Q. J. R. Meteorol.*
993 *Soc.*, **141**, 987–1003.

994 **LIST OF FIGURES**

995 **Fig. 1.** The composite evolution of the NAM index for (a) all SSWs calculated in the entire ensemble of integrations; (b) DW-propagating SSWs calculated using the Karpechko et al. (2017) criteria (see section c); (c) same as (b) except for NDW-propagating SSW events; and (d) the composite difference between the DW- and NDW-propagating events (DW-NDW). The units are in standard deviations. The thick black line in (d) represents statistical significance at the 95% level. 49

1001 **Fig. 2.** Same as figure 1 except for the anomalous vertical component of the Eliassen-Palm flux, $F^{(z)}$ (see text), averaged over the latitude band of 45-75°N and filtered for planetary waves 1 and 2. Note that $F^{(z)}$ has been scaled by the climatological standard deviation at each level so that the contours have units of standard-deviation units. Certain positive-valued contours have been added to aid in the discussion. The dashed vertical lines represent the start and end of the different lag stages used throughout the remainder of the manuscript (see text). The dashed line corresponding to zero lag has a double thickness for clarity. The thick black line in (d) represents statistical significance at the 95% level. 50

1009 **Fig. 3.** Geopotential height Z anomalies (shading; units m) at 700 hPa, averaged over the (top row) PC stage, (middle row) ONS stage and (bottom row) REC stage, and composited over: (left column) DW events; (middle column) NDW events; (right column) DW-NDW differences. Green contours show the Nov-Feb climatology calculated as the average over all of the 40 experiments with a contour interval of 25 m starting at 15 m. The thick black line is as in figure 1. 51

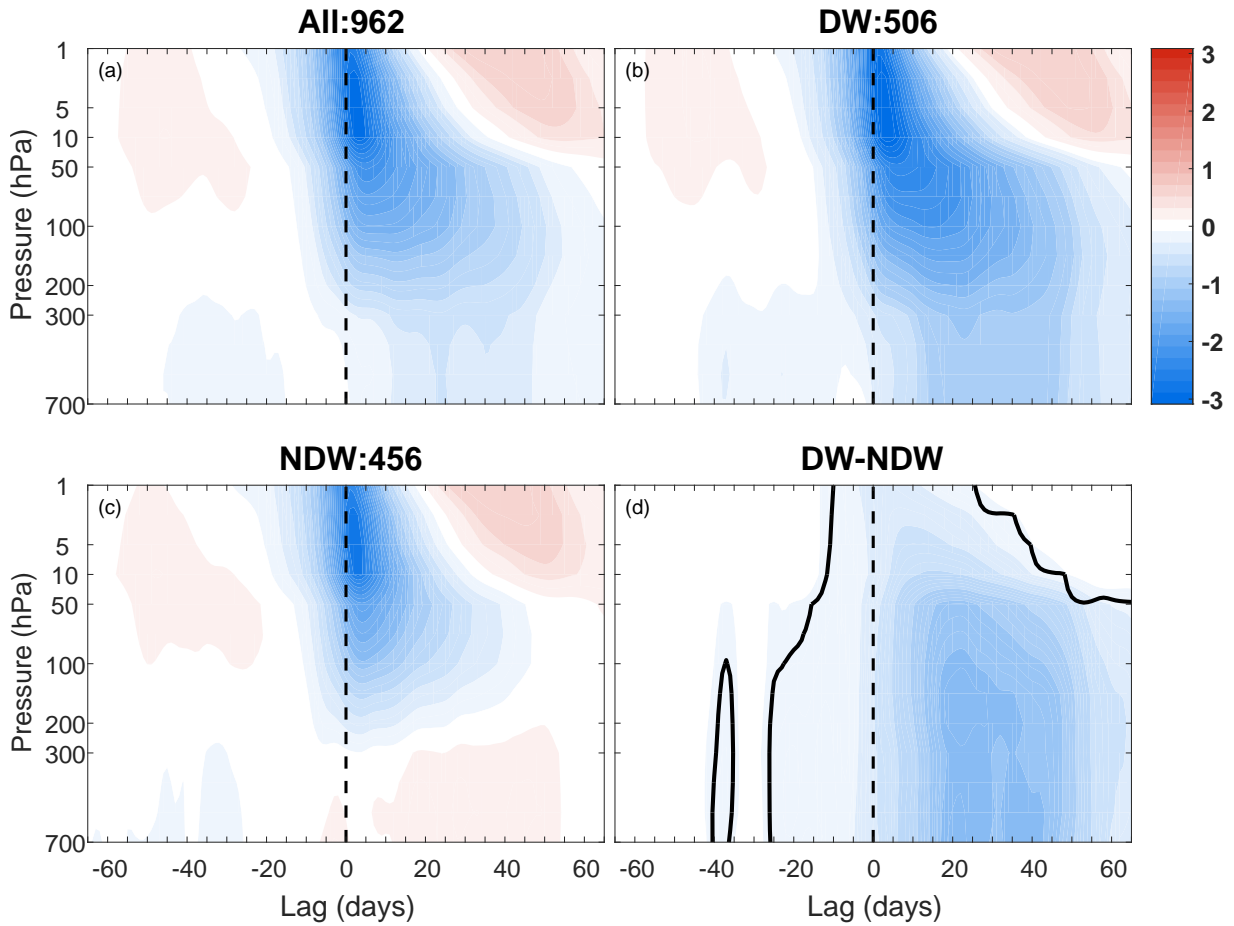
1015 **Fig. 4.** Composites of $F^{(z)}$ (filtered for waves 1-2 and standardised as in figure 2; top), NAM (middle) and Z (bottom) stratified according to the strength of $F^{(z)}$ at lags -15 to -1 at 500 hPa. (a; d; g) show the $F^{(z)}$, NAM and Z for the half of SSWs with the smallest $F^{(z)}$ anomalies (SSW_{small}), whereas (b; e; h) shows the half of SSWs with the largest $F^{(z)}$ anomalies (SSW_{large}). (c; f; i) show the corresponding $SSW_{large} - SSW_{small}$ differences. Thick black lines in (c; f; i) as in figure 1. Green contours in bottom row are as in figure 3. 52

1021 **Fig. 5.** Same as figure 3 except for the longitude-height cross-sections of Z' (i.e., deviation from the zonal-mean) averaged over the latitude band 50-60°N. The units are in m . Thin black contours show the Nov-Feb climatology calculated as the average over all of the 40 experiments with contours at -650,-550,....,550,650 m. 53

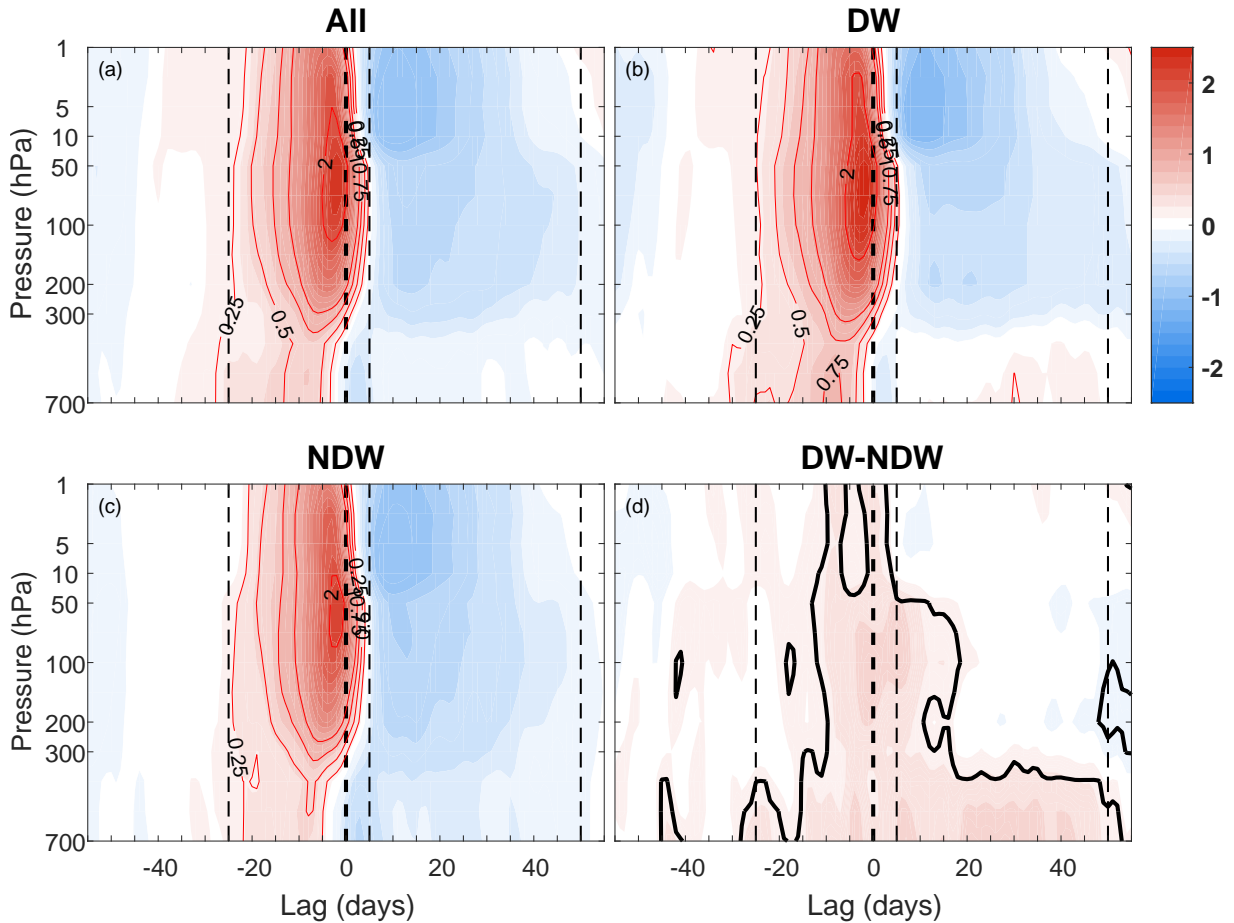
1025 **Fig. 6.** Confidence intervals for the difference (DW-NDW) of (a) the NAM index averaged over lags -25 to -1 and at 700 hPa, (b) $F^{(z)}$ anomalies at 700 hPa filtered for waves 1-2 and area-averaged over 45-75°N, and (c) Z anomalies at 700 hPa averaged over 50-80°N, 30-90°E and over lags -25 to -1. The confidence intervals are estimated using a Monte Carlo simulation of 100,000 repetitions for different sample sizes ranging from 10 to 455. The red, green and blue curves represent the 90%, 95% and 99% confidence intervals, and the respective coloured vertical dotted lines represent the sample size for which the upper bound crosses zero (indicated by the dashed black line). The dotted black line represents the overall DW-NDW composite over all DW and NDW events, as shown in figures 1- 3, respectively. 54

1034 **Fig. 7.** NAM index composited for (a) Tneg, and (b) Tpos tropospheric NAM events which have been randomly selected (see text for more details) independent of a SSW influence above. (c) shows the Tneg-Tpos composite difference with the the thick black contour the same as in figure 1. 55

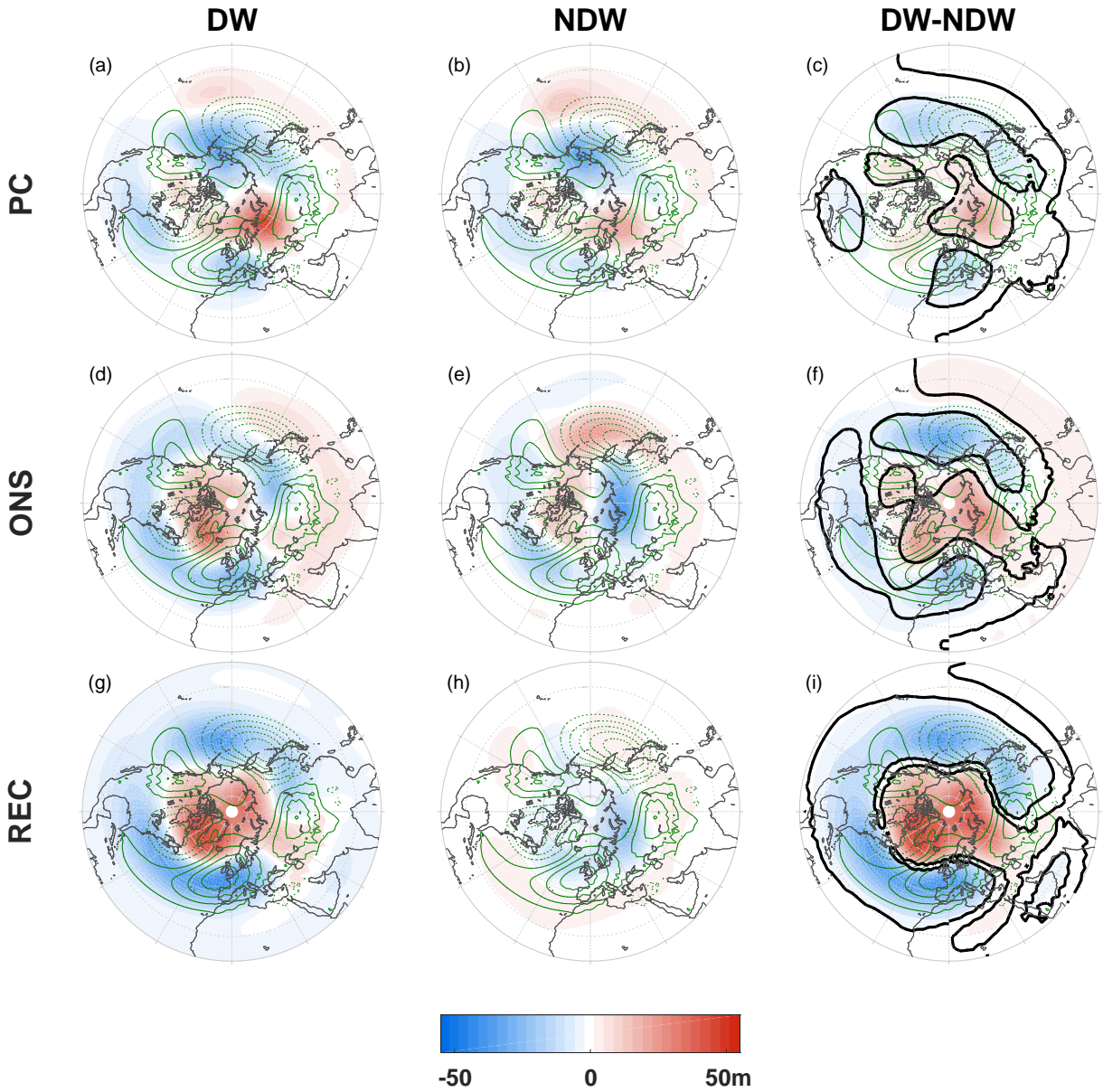
1038	Fig. 8.	Z anomalies at 700 hPa averaged over the PC stage (lags -25 to -1) for the (a) DW SSWs composite, (b) Tneg events composite, (c) DW-Tneg difference, (d) NDW SSWs composite, (e) Tpos events composite, and (f) NDW-Tpos difference. See figure 3 for details on the shading and different contours. Note that panels (a) and (d) are repeated from panels (a) and (b) in figure 3.	56
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1043	Fig. 9.	Scatter plots of standardised $F^{(z)}$ (filtered for wave 1) at (a) 100 hPa, (b) 300 hPa, and (c) 700 hPa, averaged over lags -15 to -1, against the NAM index at 10 hPa averaged over lags +1 to +10. Blue (green) diamonds, lines and squares represent, respectively, individual DW (NDW) SSW events, the corresponding lines of best fit, and the overall composite averages. The rDW (pDW), rNDW (pNDW) and r (p) represent the correlation coefficients and p-values for the DW events, NDW events, and total, respectively. The values in the top left show the numbers of DW and NDW SSWs which are preceded by such extreme wave activity averaged over lags -15 to -1.	57
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1051	Fig. 10.	Line plots of (a) the percentage of SSWs which are preceded by extreme (>2 standard deviations) $F^{(z)}$ at each level for wave-1 (green), wave-2 (red), and waves 1-2 combined (blue), and (b) the percentages of SSWs which are preceded by extreme wave activity at each level which go on to be DW (magenta) or NDW (cyan) propagating. Inset into (a) are the numbers and percentages of SSWs (rounded to the nearest percent) preceded by lower-tropospheric wave events (LTWEs; 700 hPa) to be compared with Birner and Albers (2017), as well as the numbers of extreme wave-activity events which are followed by a SSW event.	58
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1058	Fig. 11.	Composite evolution of the NAM index divided into displacements (left column) and splits (middle column) and subdivided further into the total (top row), DW-propagating (middle row) and NDW-propagating (bottom row). The right column shows the Disp-Split (top), DW-NDW displacements (middle), and DW-NDW splits (DW-NDW). See figure 1 for further details on shading and different contours.	59
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1063	Fig. 12.	As in top row of figure 3 except for Z at 700 hPa for (top) displacement SSWs and (bottom) split SSWs. Note that the green contours show the climatological Z filtered only for (top) wave-1 and (bottom) wave-2 and with a contour interval of 10 m.	60
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1066	Fig. 13.	Height-time plot of F_z averaged over 45-75°N, for the displacement SSWs composited over (left column) DW events, (middle) NDW events and (right) DW-NDW difference. Top row shows F_z for wave-1 and bottom row shows F_z for wave-2. Thick black contour in the difference plots represent statistical significance at the 95% level.	61
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1070	Fig. 14.	Same as figure 13 except for split SSW events.	62



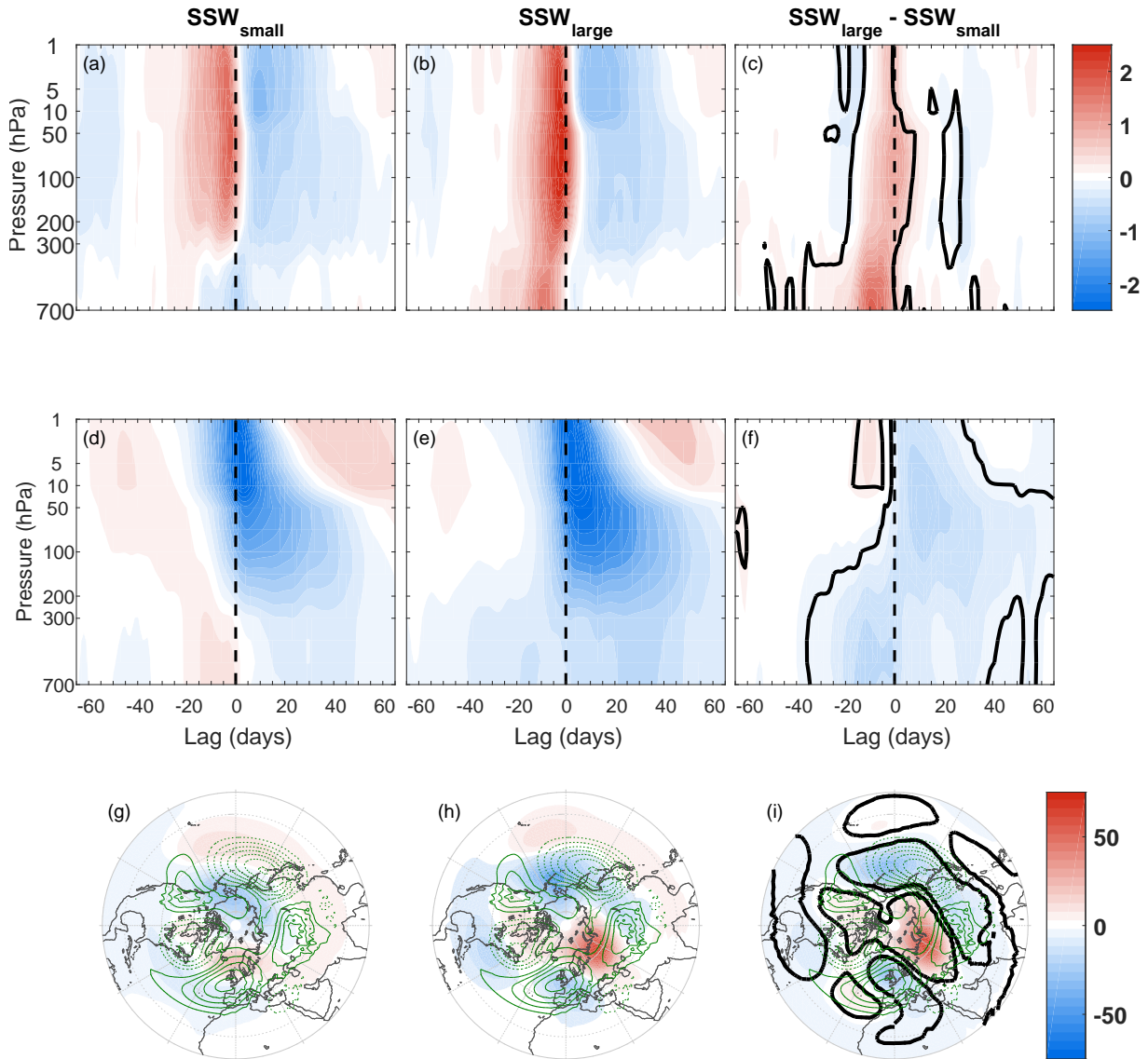
1071 FIG. 1. The composite evolution of the NAM index for (a) all SSWs calculated in the entire ensemble of
 1072 integrations; (b) DW-propagating SSWs calculated using the Karpechko et al. (2017) criteria (see section c);
 1073 (c) same as (b) except for NDW-propagating SSW events; and (d) the composite difference between the DW-
 1074 and NDW-propagating events (DW-NDW). The units are in standard deviations. The thick black line in (d)
 1075 represents statistical significance at the 95% level.



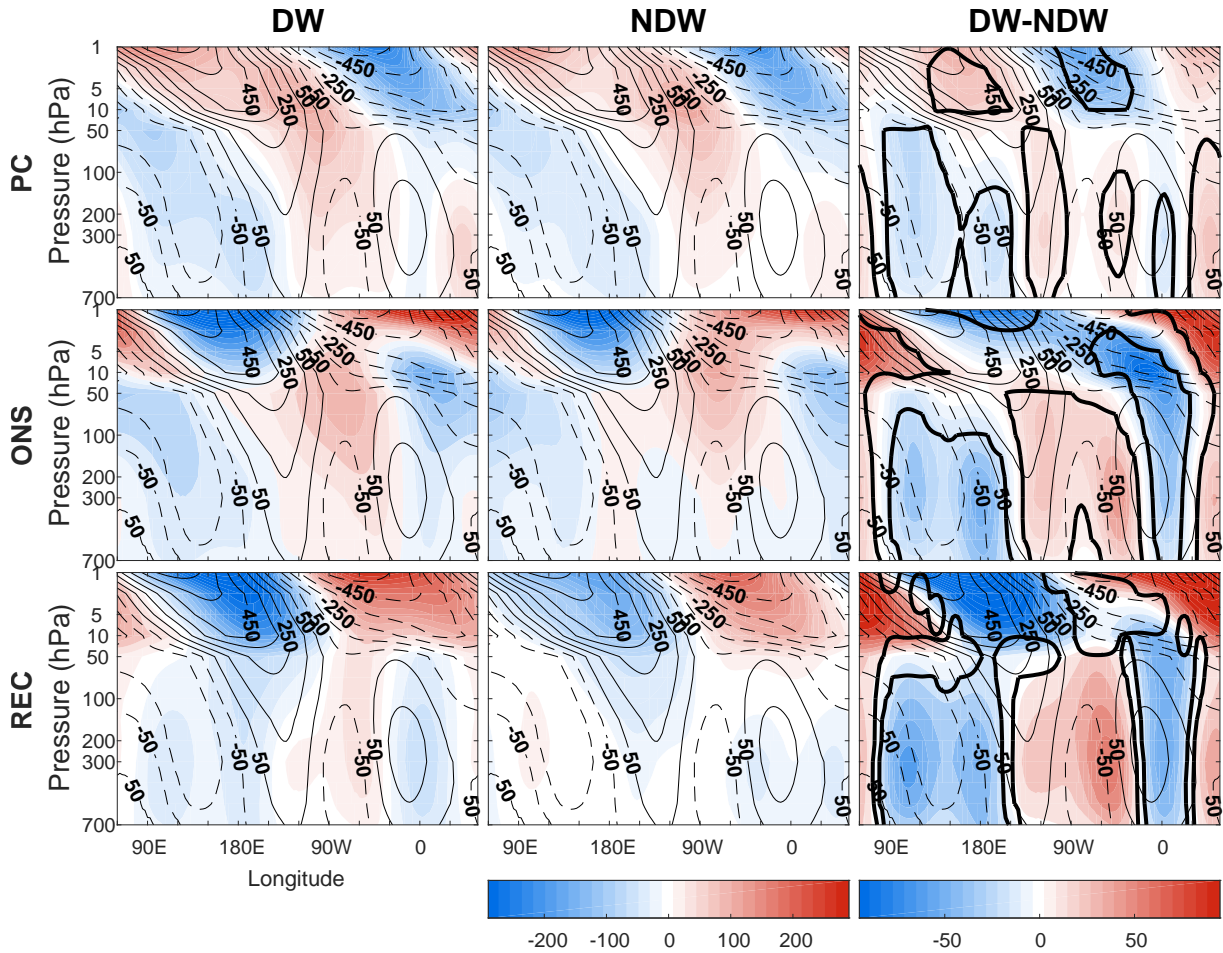
1076 FIG. 2. Same as figure 1 except for the anomalous vertical component of the Eliassen-Palm flux, $F^{(z)}$ (see
 1077 text), averaged over the latitude band of 45-75°N and filtered for planetary waves 1 and 2. Note that $F^{(z)}$ has
 1078 been scaled by the climatological standard deviation at each level so that the contours have units of standard-
 1079 deviation units. Certain positive-valued contours have been added to aid in the discussion. The dashed vertical
 1080 lines represent the start and end of the different lag stages used throughout the remainder of the manuscript (see
 1081 text). The dashed line corresponding to zero lag has a double thickness for clarity. The thick black line in (d)
 1082 represents statistical significance at the 95% level.



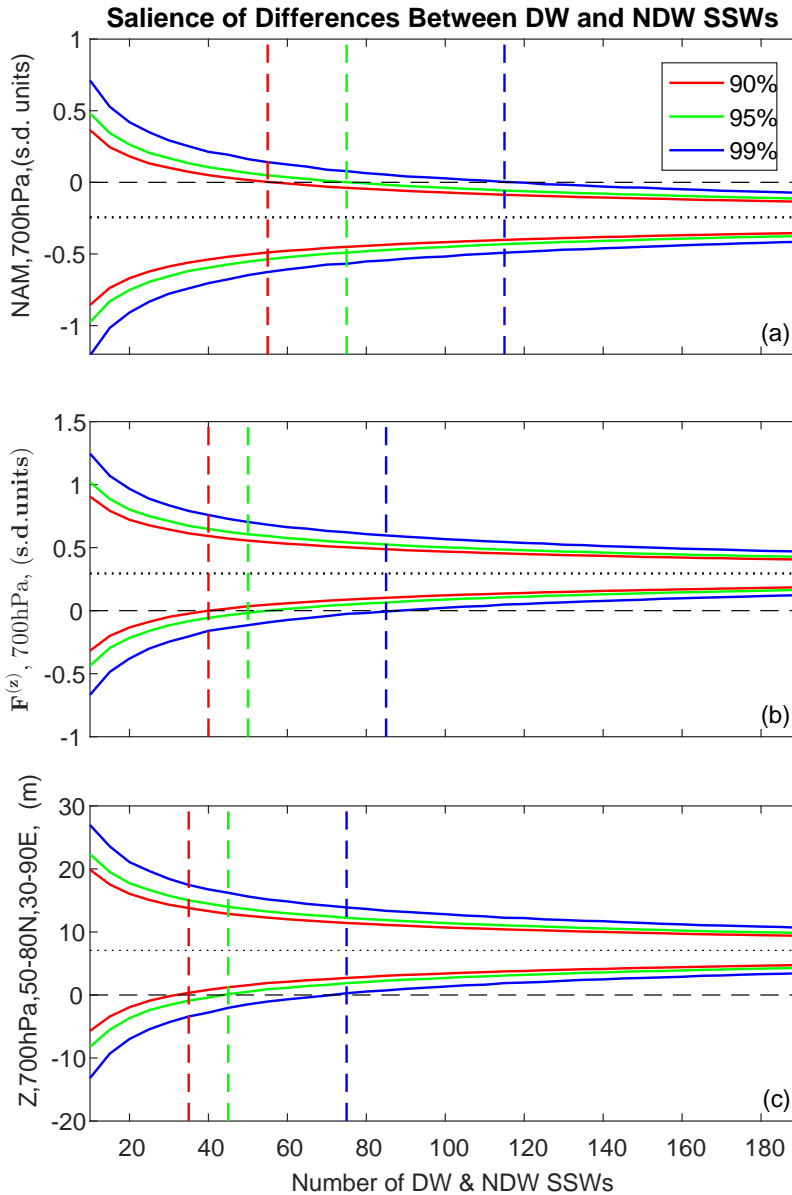
1083 FIG. 3. Geopotential height Z anomalies (shading; units m) at 700 hPa, averaged over the (top row) PC stage,
 1084 (middle row) ONS stage and (bottom row) REC stage, and composited over: (left column) DW events; (middle
 1085 column) NDW events; (right column) DW-NDW differences. Green contours show the Nov-Feb climatology
 1086 calculated as the average over all of the 40 experiments with a contour interval of 25 m starting at 15 m. The
 1087 thick black line is as in figure 1.



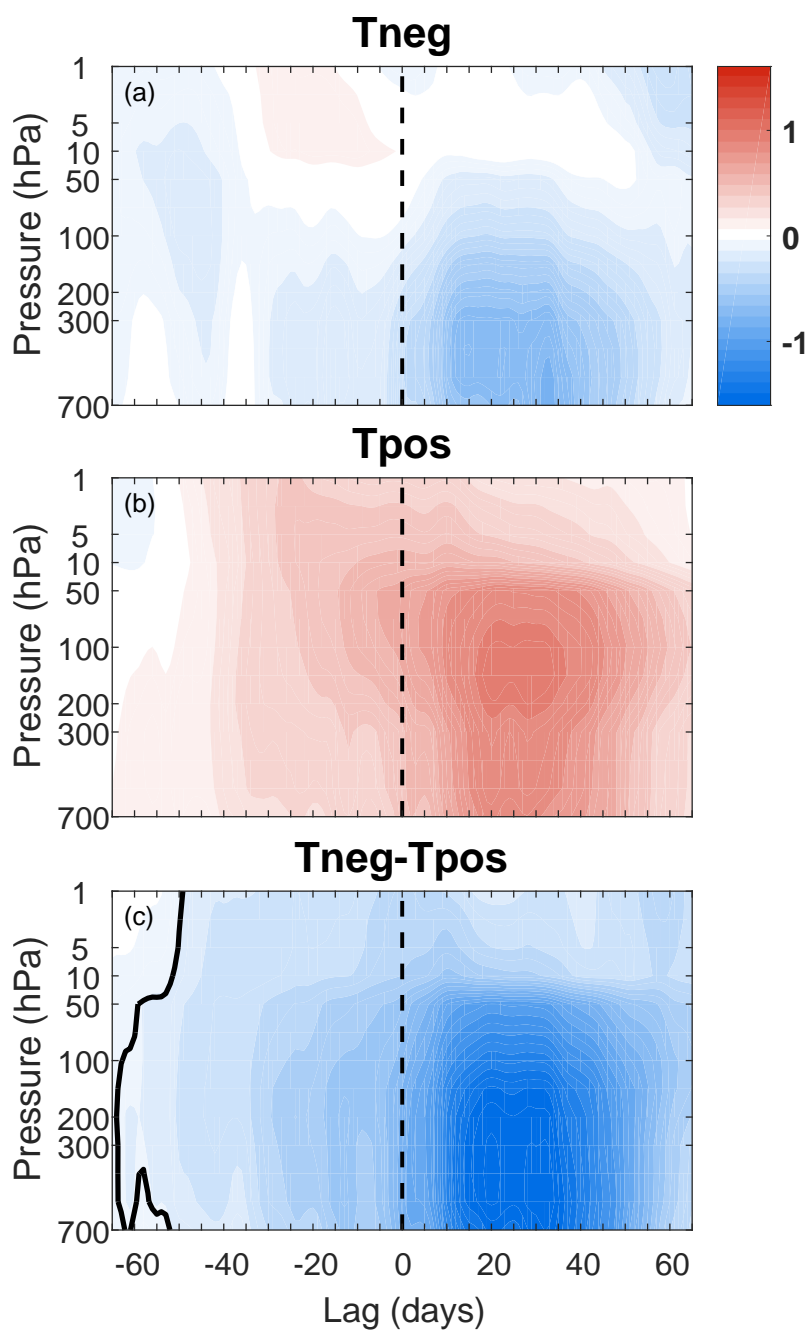
1088 FIG. 4. Composites of $F^{(z)}$ (filtered for waves 1-2 and standardised as in figure 2; top), NAM (middle) and Z
 1089 (bottom) stratified according to the strength of $F^{(z)}$ at lags -15 to -1 at 500 hPa. (a; d; g) show the $F^{(z)}$, NAM and
 1090 Z for the half of SSWs with the smallest $F^{(z)}$ anomalies (SSW_{small}), whereas (b; e; h) shows the half of SSWs
 1091 with the largest $F^{(z)}$ anomalies (SSW_{large}). (c; f; i) show the corresponding $SSW_{\text{large}} - SSW_{\text{small}}$ differences.
 1092 Thick black lines in (c; f; i) as in figure 1. Green contours in bottom row are as in figure 3.



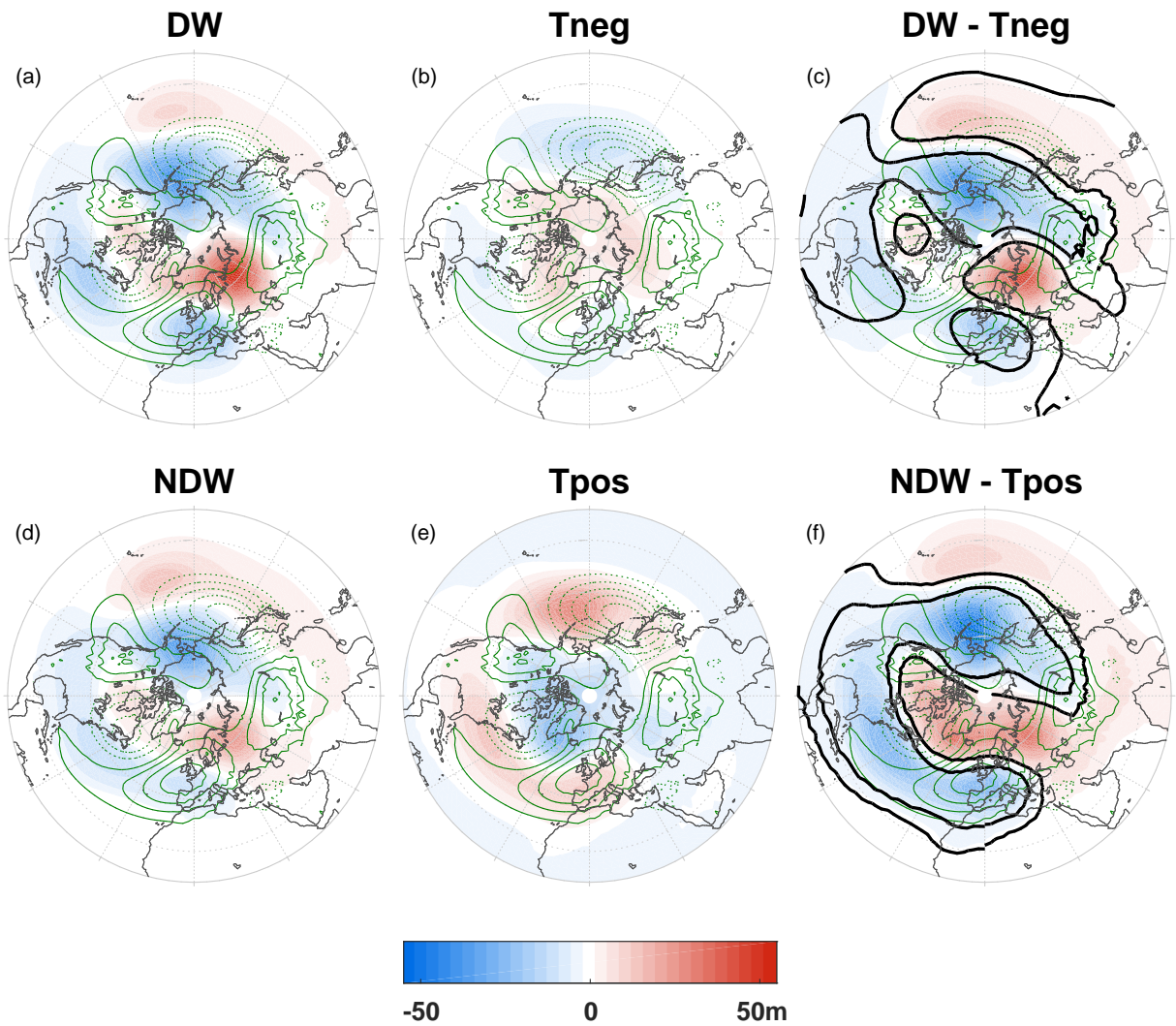
1093 FIG. 5. Same as figure 3 except for the longitude-height cross-sections of Z' (i.e., deviation from the zonal-
 1094 mean) averaged over the latitude band 50-60°N. The units are in m . Thin black contours show the Nov-Feb
 1095 climatology calculated as the average over all of the 40 experiments with contours at -650,-550,...,550,650 m .



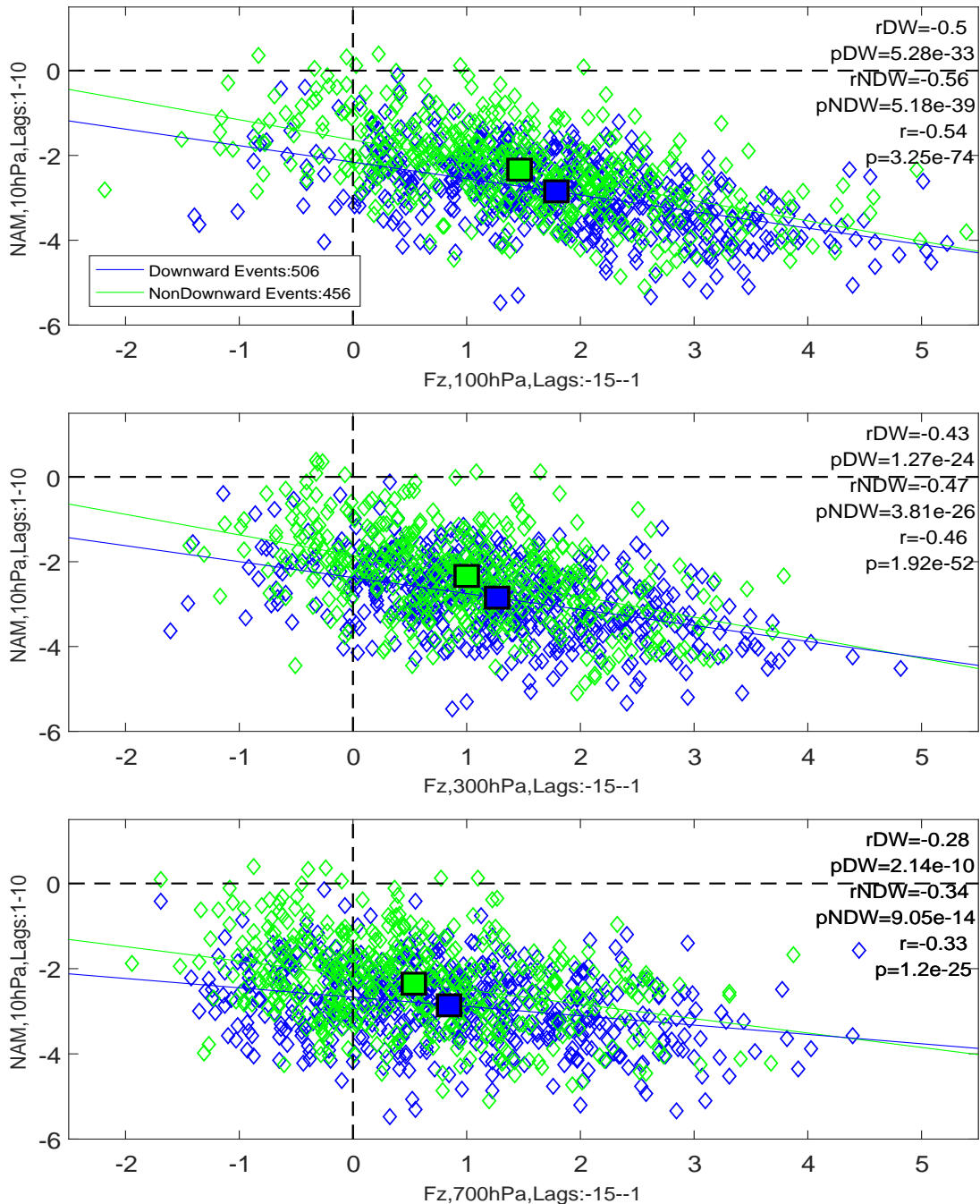
1096 FIG. 6. Confidence intervals for the difference (DW-NDW) of (a) the NAM index averaged over lags -25 to
 1097 -1 and at 700 hPa, (b) $F^{(z)}$ anomalies at 700 hPa filtered for waves 1-2 and area-averaged over 45-75°N, and (c)
 1098 Z anomalies at 700 hPa averaged over 50-80°N, 30-90°E and over lags -25 to -1. The confidence intervals are
 1099 estimated using a Monte Carlo simulation of 100,000 repetitions for different sample sizes ranging from 10 to
 1100 455. The red, green and blue curves represent the 90%, 95% and 99% confidence intervals, and the respective
 1101 coloured vertical dotted lines represent the sample size for which the upper bound crosses zero (indicated by the
 1102 dashed black line). The dotted black line represents the overall DW-NDW composite over all DW and NDW
 1103 events, as shown in figures 1- 3, respectively.



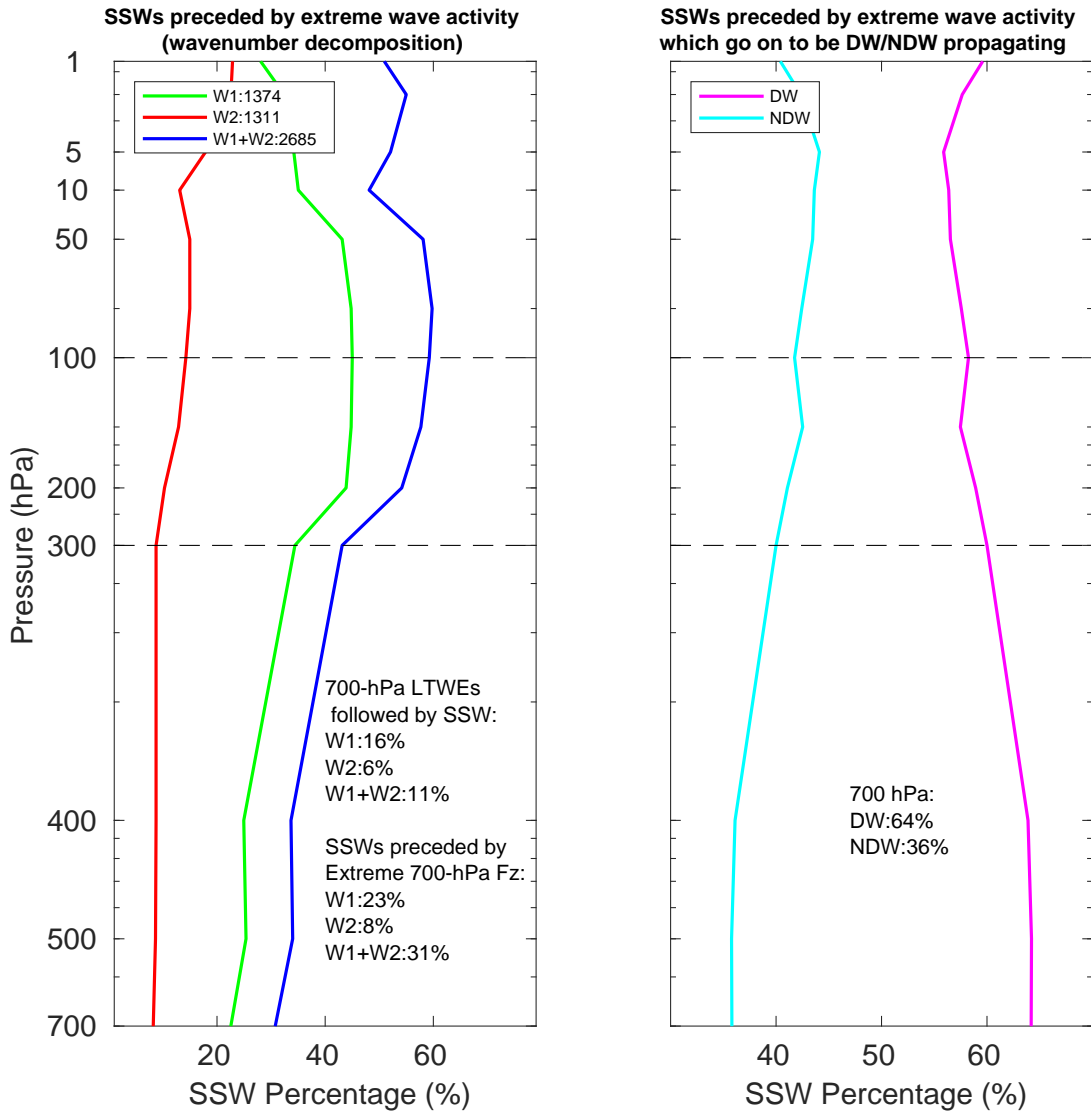
1104 FIG. 7. NAM index composited for (a) Tneg, and (b) Tpos tropospheric NAM events which have been
 1105 randomly selected (see text for more details) independent of a SSW influence above. (c) shows the Tneg-Tpos
 1106 composite difference with the the thick black contour the same as in figure 1.



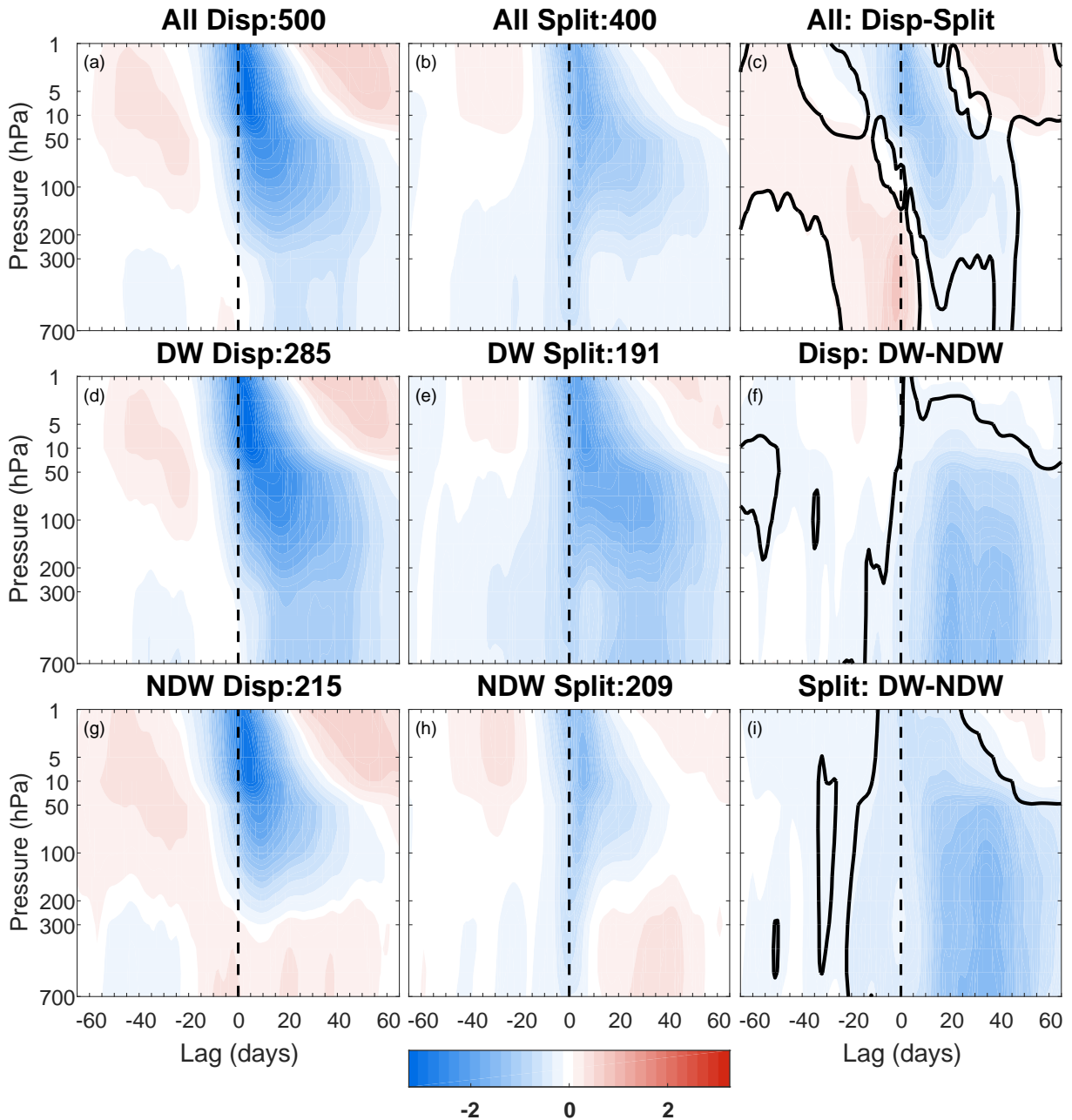
1107 FIG. 8. Z anomalies at 700 hPa averaged over the PC stage (lags -25 to -1) for the (a) DW SSWs composite,
 1108 (b) Tneg events composite, (c) DW-Tneg difference, (d) NDW SSWs composite, (e) Tpos events composite, and
 1109 (f) NDW-Tpos difference. See figure 3 for details on the shading and different contours. Note that panels (a)
 1110 and (d) are repeated from panels (a) and (b) in figure 3.



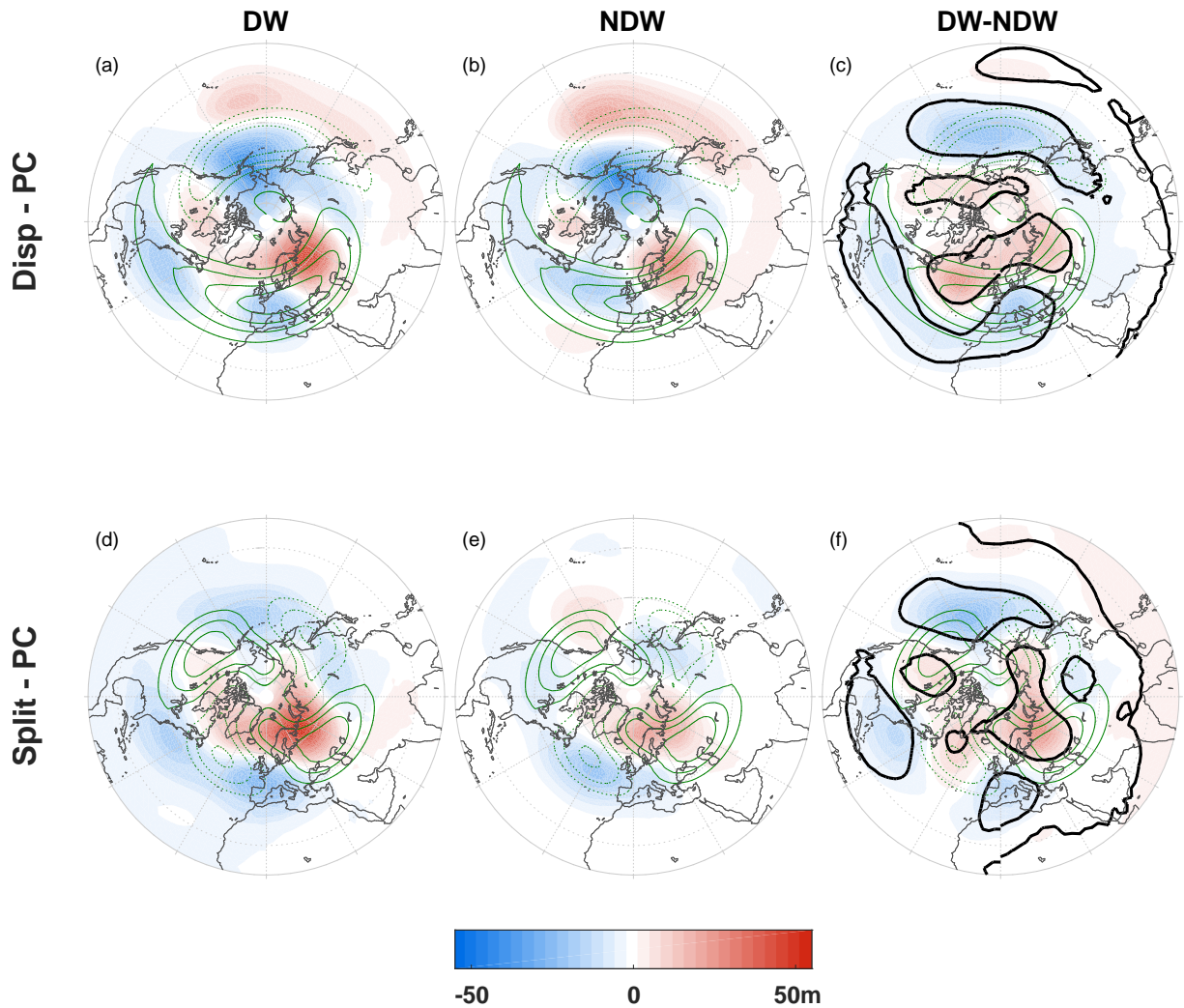
1111 FIG. 9. Scatter plots of standardised $F^{(z)}$ (filtered for wave 1) at (a) 100 hPa, (b) 300 hPa, and (c) 700 hPa,
 1112 averaged over lags -15 to -1, against the NAM index at 10 hPa averaged over lags +1 to +10. Blue (green)
 1113 diamonds, lines and squares represent, respectively, individual DW (NDW) SSW events, the corresponding
 1114 lines of best fit, and the overall composite averages. The rDW (pDW), rNDW (pNDW) and r (p) represent the
 1115 correlation coefficients and p-values for the DW events, NDW events, and total, respectively. The values in the
 1116 top left show the numbers of DW and NDW SSWs which are preceded by such extreme wave activity averaged
 1117 over lags -15 to -1.



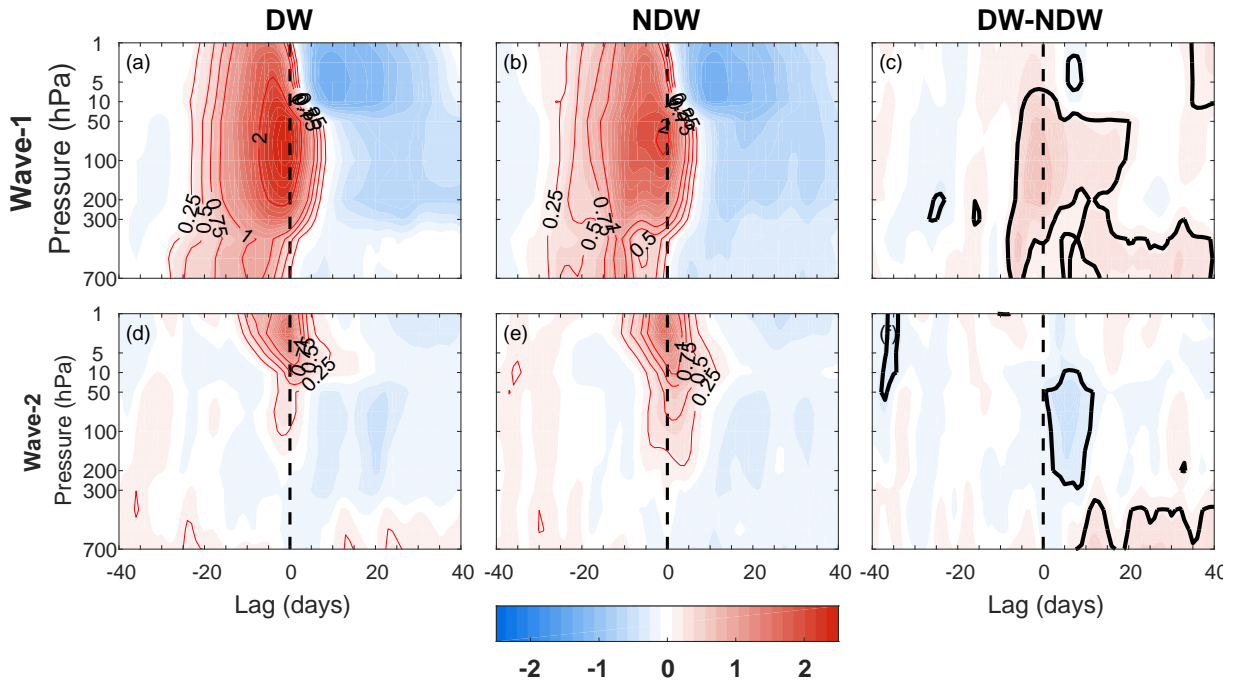
1118 FIG. 10. Line plots of (a) the percentage of SSWs which are preceded by extreme (>2 standard deviations)
1119 $F^{(z)}$ at each level for wave-1 (green), wave-2 (red), and waves 1-2 combined (blue), and (b) the percentages of
1120 SSWs which are preceded by extreme wave activity at each level which go on to be DW (magenta) or NDW
1121 (cyan) propagating. Inset into (a) are the numbers and percentages of SSWs (rounded to the nearest percent)
1122 preceded by lower-tropospheric wave events (LTWEs; 700 hPa) to be compared with Birner and Albers (2017),
1123 as well as the numbers of extreme wave-activity events which are followed by a SSW event.



1124 FIG. 11. Composite evolution of the NAM index divided into displacements (left column) and splits (middle
 1125 column) and subdivided further into the total (top row), DW-propagating (middle row) and NDW-propagating
 1126 (bottom row). The right column shows the Disp-Split (top), DW-NDW displacements (middle), and DW-NDW
 1127 splits (DW-NDW). See figure 1 for further details on shading and different contours.



1128 FIG. 12. As in top row of figure 3 except for Z at 700 hPa for (top) displacement SSWs and (bottom) split
 1129 SSWs. Note that the green contours show the climatological Z filtered only for (top) wave-1 and (bottom)
 1130 wave-2 and with a contour interval of 10 m.



1131 FIG. 13. Height-time plot of F_z averaged over $45\text{-}75^\circ\text{N}$, for the displacement SSWs composited over (left
 1132 column) DW events, (middle) NDW events and (right) DW-NDW difference. Top row shows F_z for wave-1 and
 1133 bottom row shows F_z for wave-2. Thick black contour in the difference plots represent statistical significance at
 1134 the 95% level.

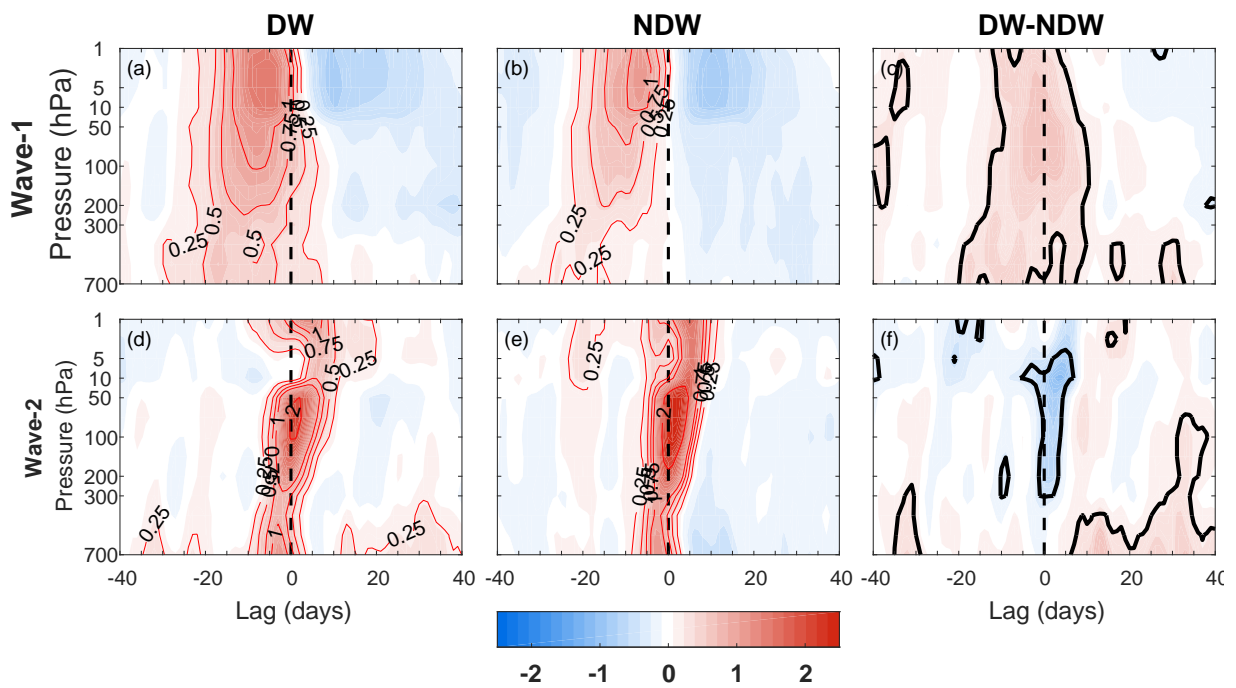


FIG. 14. Same as figure 13 except for split SSW events.

1 **Supplement to: The Downward Influence of Sudden Stratospheric**

2 **Warmings: Association with Tropospheric Precursors**

3 Ian White*

4 *The Hebrew University of Jerusalem, Institute of Earth Sciences, Edmond J. Safra Campus, Givat*
5 *Ram, Jerusalem, Israel*

6 Chaim I. Garfinkel

7 *The Hebrew University of Jerusalem, Institute of Earth Sciences, Edmond J. Safra Campus, Givat*
8 *Ram, Jerusalem, Israel*

9 Edwin P. Gerber

10 *Courant Institute of Mathematical Sciences, New York University, New York, USA*

11 Martin Jucker

12 *School of Earth Sciences, The University of Melbourne, Parkville, Australia*

13 Valentina Aquila

14 *American University, Dept of Environmental Science, Washington, DC, USA*

15 Luke D. Oman

16 *NASA Goddard Space Flight Center, Greenbelt, Maryland, USA*

¹⁷ **Corresponding author address:* Ian White, The Hebrew University of Jerusalem, Institute of Earth
¹⁸ Sciences, Edmond J. Safra Campus, Givat Ram, Jerusalem, Israel.
¹⁹ E-mail: ian.white@mail.huji.ac.il

ABSTRACT

20

21 **1. Other Downward-Propagating Definitions**

22 We here provide a more detailed description of the three extra definitions of downward (DW)
23 propagating SSWs used in this study of which the numbers are summarised in table 1 in the main
24 text; one by Runde et al. (2016) and two by Jucker (2016):

25 **Runde et al. (2016)**

26 They proposed a more restrictive definition than that by Karpechko et al. (2017). In particular,
27 the NAM index has to be more negative than -1.5 standard deviations at every level below 10
28 hPa down to 850 hPa for at least one day in the succeeding 70 days (although we chose this
29 window) after the onset date. Additionally, the date of the first exceedance of the threshold at
30 each level must be after (or occur simultaneously to) the first exceedance at the level above.
31 If this is not satisfied then the end date of the exceedance at a given level must occur after (or
32 again simultaneously to) the end date at the level above. Further, the start lag of the threshold
33 exceedance at a given level must be within 30 days of the end date of threshold exceedance
34 at the level above, to try and ensure that the anomalies at each level are connected (personal
35 communication). Overall this ensures that there is a seemingly DW propagation from the middle
36 stratosphere to the lower troposphere.

37

38 **Jucker (2016)**

39 The two proposed definitions by Jucker (2016) are referred to here and in the main text as the
40 absolute-criterion and relative-criterion definitions. The absolute-criterion definition simply
41 demands that the NAM index averaged over lags +10 to +40 be smaller than -0.6. We note, as they
42 do, that our results are insensitive to changes in this window, as well as changes in the threshold
43 value. On the other hand, the relative-criterion definition demands that the relative change of the

44 NAM index at 500 hPa between positive lags (averaged over lags +1 to +80) and negative (aver-
45 aged over lags -80 to -1) must be smaller than -0.1. We note again, that the results are not sensitive
46 to the thresholds in this definition, aside from the fact that the averaging periods used influence
47 the width of the positive and negative anomalies either side of the onset date in the composite plots.

48
49 One thing to be mindful of when identifying a given SSW as DW-propagating is to determine
50 if the tropospheric NAM anomalies are actually attributable to those in the stratosphere. More
51 specifically, the negative tropospheric NAM at positive lags could be due to either the stratospheric
52 anomaly propagating DW, or, due to the persistence of a negative tropospheric NAM prior to the
53 onset. Indeed, it could also be a combination of the two, or even the negative tropospheric NAM
54 at positive lags spontaneously developing, unrelated to the stratosphere. Of course, to distinguish
55 between all of these is very difficult, but from our sensitivity tests, the definitions by Karpechko
56 et al. (2017) and Runde et al. (2016) go some way towards ensuring this, with particular emphasis
57 on the latter which demands an apparent systematic DW propagation from the middle stratosphere
58 to near the surface. Note that the Karpechko and Runde definitions yield quantitatively similar
59 results, and because the former gives a larger compendium of DW SSWs (see table 1 in the main
60 text), we choose to mostly utilise the definition by Karpechko et al. (2017) there, unless explicitly
61 stated otherwise.

62 63 **2. Robustness of Precursors to DW Definition**

64 We now test the robustness of the zonal-mean NAM precursors using each of the DW def-
65 initions introduced above. Figure 1 shows the NAM index at 500 hPa for the DW definition
66 of Karpechko et al. (2017) (red line; see also figure 1 in main text), Runde et al. (2016) (blue

67 line), and the absolute- and relative-criterion definitions of Jucker (2016) (green and black lines,
68 respectively). We first note that at positive lags, all definitions show negative NAM for DW events
69 by construction, although with differing magnitudes depending on the thresholds used in the
70 individual definitions. At negative lags, the Karpechko, Runde and absolute-criterion definitions
71 give quantitatively similar results to one another, with the DW composite showing negative NAM
72 values prior to lag zero, and the NDW composite showing positive values from approximately lag
73 -20 to 0 and negative values beforehand. This gives differences that are therefore negative and
74 statistically significant extending back to approximately lag -25.

75
76 The relative-criterion definition gives drastically different results however for the DW and
77 NDW composites prior to lag zero; positive anomalies for DW events and negative anomalies for
78 NDW events, yielding positive differences prior to lag zero. The differences are antisymmetric
79 (although the negative NAM at positive lags is of larger magnitude) around the central date and
80 this is found to depend on the averaging window used to determine the DW propagation; in this
81 example we used lags -40 to -10 and lags +10 to +40 as the averaging periods. This also agrees
82 with Jucker (2016) who showed a similar composite centred on lag zero.

83
84 The differences in the NAM evolution among the four definitions can be related to the periods
85 of time used in each definition. For instance, the Karpechko, Runde, and absolute-criterion
86 definitions only use values of the NAM at positive lags, whereas the relative-criterion uses NAM
87 values at both negative and positive lags. In regards to the former three, they can be used to
88 identify possible precursor features at negative lags (and in fact, the Karpechko definition can be
89 used up until lag +7) as required for this study, as they do not force the composites at such lags.
90 In the case of the relative-criterion definition however, any precursors may be influenced by the

91 definition. For this reason, we believe that the presence of the precursors are robust but we note
92 that they are sensitive to the type of definition used.

93

94 **References**

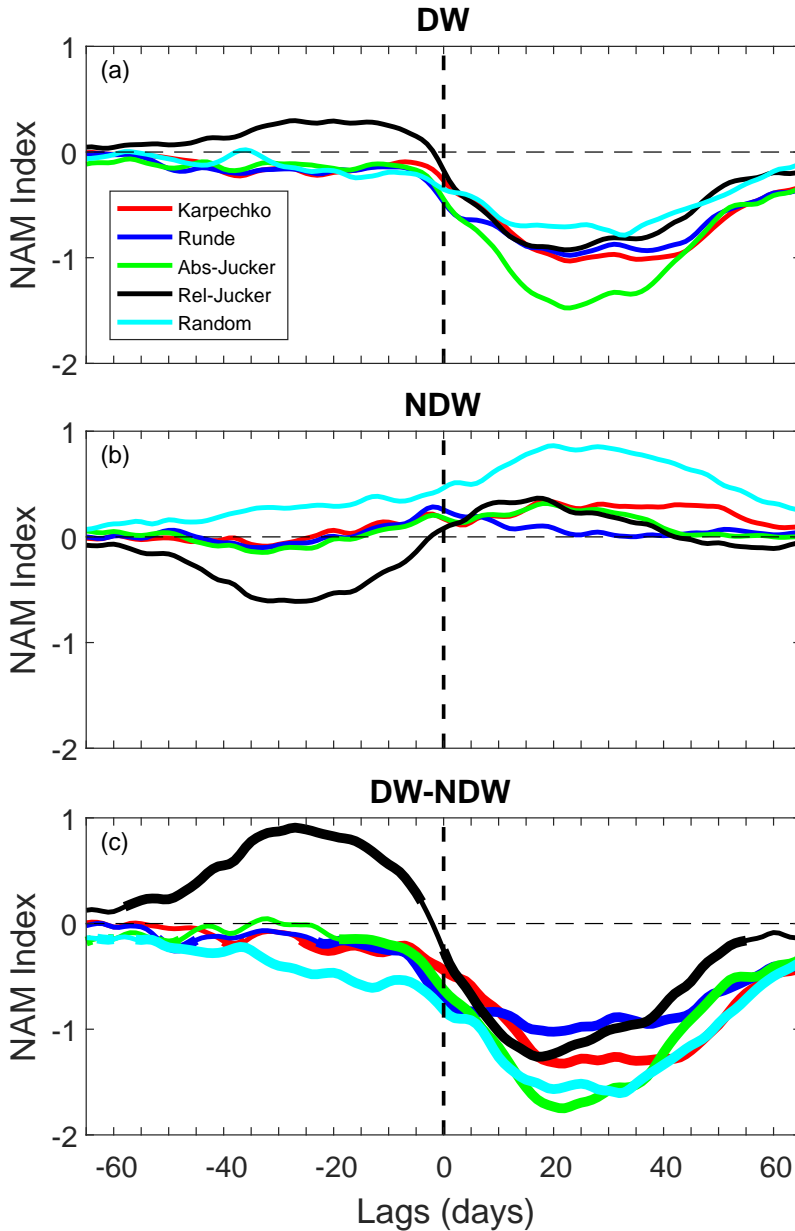
95 Jucker, M., 2016: Are sudden stratospheric warmings generic? insights from an idealized gcm. *J.*
96 *Atmos. Sci.*, **73**, 5061–5080.

97 Karpechko, A. Y., P. Hitchcock, D. H. W. Peters, and A. Schneidereit, 2017: Predictability of
98 downward propagation of major sudden stratospheric warmings. *Q. J. R. Meteorol. Soc.*, **143**,
99 1459–1470.

100 Runde, T., M. Dameris, H. Garny, and D. E. Kinnison, 2016: Classification of stratospheric ex-
101 treme events according to their downward propagation to the troposphere. *Geophys. Res. Lett.*,
102 **43**, 6665–6672.

103 **LIST OF FIGURES**

104 **Fig. 1.** NAM index at 500 hPa composited over (a) DW events, (b) NDW events, and (c) DW-
105 NDW differences, for the four DW definitions introduced previously: red, dark blue, green
106 and black lines represent the NAM index for the Karpechko et al. (2017) definition, Runde
107 et al. (2016) definition, and absolute- and relative-criterion definitions of Jucker (2016),
108 respectively. There is also an additional cyan line representing the NAM index found using
109 a random selection of tropospheric NAM events (see main text). The thick lines represent
110 statistical significance at the 95% level. 9



111 FIG. 1. NAM index at 500 hPa composited over (a) DW events, (b) NDW events, and (c) DW-NDW differ-
 112 ences, for the four DW definitions introduced previously: red, dark blue, green and black lines represent the
 113 NAM index for the Karpechko et al. (2017) definition, Runde et al. (2016) definition, and absolute- and relative-
 114 criterion definitions of Jucker (2016), respectively. There is also an additional cyan line representing the NAM
 115 index found using a random selection of tropospheric NAM events (see main text). The thick lines represent
 116 statistical significance at the 95% level.