Linking warm Arctic winters, Rossby waves and Cold Spells: an idealized numerical study

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ABSTRACT

12 The changes of midlatitude Rossby waves and cold extreme temperature
13 events (cold spells) during warm Arctic winters are analysed using a dry three-
14 level quasi-geostrophic model on the sphere. Two long-term simulations are
15 compared: the first run has the observed wintertime climatology, while the
16 second run includes the composite of the global anomalies associated with the
17 six hottest Arctic winters. A spectral analysis shows a large increase in wave
18 amplitude for near-zero and westward phase speeds and a more moderate de-
19 crease for high eastward phase speeds. The increase in low-frequency vari-
20 ability (periods greater than a week) associated with the power shift to slower
21 waves is largely responsible for an increase in mid-latitude long-lasting cold
22 spells. In mid-latitude regions, in presence of a mean warming, that increase
23 in low-frequency variance compensates the increase of the mean temperature,
24 resulting at places in a frequency of cold spells that remains by and large
25 unaltered. In presence of mean cooling, both the increase in variance and
26 the decrease in the mean temperature participate in an increased frequency of
27 cold spells. Sensitivity experiments show that the power shift to slower waves
28 is mainly due to the tropical anomalies that developed during those particu-
29 lar winters and less importantly to changes in the background flow at higher
30 latitudes associated with the Arctic Amplification pattern.

31 **1. Introduction**

32 Observations show that climate warming is not homogeneous over the globe, and *Arctic Ampli-*
33 *fication* (AA) refers to the accelerated rate of warming of the Arctic near the surface compared to
34 tropics and mid-latitudes (Serreze et al. 2009). AA shows a seasonality, with a maximum during
35 late fall and winter (Lu and Cai 2009), it can be found in historical observations, in projection of
36 future climate (Pithan and Mauritsen 2014) and even in paleoclimatic studies (Serreze and Barry
37 2011), with a consistent relationship with the concentration of greenhouse gases (Fletcher et al.
38 2018).

39 Several - non exclusive - mechanisms of AA have been identified in the literature (see Cohen
40 et al. 2018, for a recent review). The best-known mechanism is the positive albedo feedback asso-
41 ciated with sea-ice reduction and snow melting (Screen and Simmonds 2010). Another category
42 of mechanisms of AA involves an increased downward infrared radiation due to changes in cloud
43 cover and water vapor (Graversen and Wang 2009). Intrusions of moisture from lower latitudes
44 might largely be the cause of such changes (Woods and Caballero 2016; Gong et al. 2017; Lee
45 et al. 2017), due to synoptic dynamics in the Arctic (Inoue and Hori 2011; Rinke et al. 2017; Mes-
46 sori et al. 2018) or in lower latitudes (Binder et al. 2017) or due to planetary waves initiated in the
47 tropics (Lee et al. 2011).

48 The possible influence of AA on midlatitude weather has been the subject of an intense debate
49 during the last decade. In boreal summer, some studies advanced a key role played by Arctic
50 warming in increasing the occurrence of heat waves, droughts and flooding events via Rossby
51 waves quasi-resonance and changes in jet behavior (Coumou et al. 2014, 2015; Mann et al. 2018).
52 Di Capua and Coumou (2016) found the most robust changes and a significant positive trend in
53 waviness during fall. However, most of the debate is focused on boreal winter, the AA signal

54 is in fact strongest in winter, and there have been concomitant trends of Arctic warming and
55 Northern Hemisphere continents cooling since the 1990s (Cohen et al. 2012b). There have also
56 been specific seasons marked by extreme AA, associated with records in sea ice reduction in
57 autumn followed by extreme cooling over the continents in winter: this was the case of the 2005-
58 2006 winter over the Eurasian continent (Petoukhov and Semenov 2010), 2013-2014 winter over
59 North America (Lee et al. 2015) and 2009-2010 winter over the whole Hemisphere (Overland
60 et al. 2011). Finally, some studies detected an increase in the number of cold spells (CS) and
61 noted some record snowfalls in North America, Europe and Asia over the recent past winters (Liu
62 et al. 2012; Francis and Vavrus 2012; Cohen et al. 2014).

63 A first category of observational and modelling studies considered a regional influence of sea ice
64 reduction onto the midlatitudes: the Barents-Kara sea ice reduction would imply cold winter over
65 central Eurasia (Honda et al. 2009; Mori et al. 2014) while East-Siberian-Chukchi-Bering sea ice
66 reduction would induce cold weather over North America (Kug et al. 2015; Lee et al. 2015). The
67 underlying mechanism is the same, involving the formation of a high anomaly, or high-latitude
68 block, over the reduced sea ice region.

69 Another category of studies provided evidences for an hemispheric impact and for the formation
70 of the negative phase of the Arctic Oscillation / North Atlantic Oscillation (AO/NAO) in response
71 to sea ice reduction and/or AA. Such an impact would explain cold continents in winter and fa-
72 vor cold periods over North America and Eurasia (Thompson and Wallace 1998; Yiou and Nogaj
73 2004). It would also explain that the decades of 1980s' and 1990s' were marked by an increase
74 in AO/NAO index (Hurrell 1995), while the early twenty first century has been characterized by
75 near-neutral and negative AO/NAO phases (Overland and Wang 2005; Cohen et al. 2012a). Many
76 studies have also emphasized the key role played by the stratosphere to explain this influence (Co-
77 hen et al. 2012a; Peings and Magnusdottir 2014; Nakamura et al. 2015; Hell et al. 2020). While

78 observational studies have emphasized this close relationship between reduced sea-ice cover and
79 the negative phase of the AO/NAO, modelling studies have not robustly found the same type of
80 phenomenon (Bader et al. 2011; Cohen et al. 2014; Sun et al. 2016). Finally, it is important to
81 note that such a negative NAO/AO response to Arctic warming, which corresponds to an equa-
82 torward shift of the midlatitude jets, can also be found in simple dry atmospheric models without
83 representation of the stratosphere and without land-sea contrasts (Butler et al. 2010; Rivière 2011).

84 Since the Arctic warming is only one feature of global warming, it is not obvious if its effect will
85 dominate over others. A strong warming is also expected to occur in the tropical upper troposphere,
86 with the effect of shifting the storm tracks and jet streams poleward (Butler et al. 2010; Rivière
87 2011). The opposite influences of Arctic and tropical warmings is sometimes referred to as the
88 "tug of war" (Barnes and Screen 2015; Shaw et al. 2016). The general effect of global warming
89 in CMIP3 and CMIP5 experiments is mainly characterized by a poleward shift of the jet streams,
90 suggesting that tropical warming influence dominates (Shaw et al. 2016). However, the ability
91 of the models to accurately simulate the spatial heterogeneities of the warming is still questioned
92 (Santer et al. 2018).

93 The hypothesis has been advanced that Arctic warming would favor large-amplitude meanders
94 of the jet and extreme weather events like CS (Liu et al. 2012; Francis and Vavrus 2012). Ac-
95 cording to Francis and Vavrus (2012), the phase speed of the waves should decrease because
96 midlatitude westerly jets decrease in intensity in association with the decreased meridional tem-
97 perature gradient; additionally, the ridges of the waves should become more elongated because of
98 the presence of a positive geopotential height anomaly in the Arctic in connection with the ampli-
99 fied warming. Both effects would induce more amplified and slower-propagating waves and finally
100 more extreme and more persistent temperature events. In particular, this would explain the recent
101 cold and snowy winters. Over Europe, CS are generally induced by the presence of a blocking

102 (Buehler et al. 2011) while over North America large-scale precursors of CS are more associated
103 with amplified slowly-propagating wave trains (Grotjahn et al. 2016; Messori et al. 2016; Harnik
104 et al. 2016). However, there is no indication of a positive trend in blocking frequency in reanalysis
105 datasets (Barnes et al. 2014; Davini and D’Andrea 2020), or when detected, they are specific to
106 some regions only (Davini et al. 2012). Furthermore, a decrease in blockings has been detected
107 in CMIP5 and CMIP6 future climate scenarios (Masato et al. 2013; Davini and D’Andrea 2020).
108 Other studies measured the intensity and the propagation speed of synoptic and planetary waves
109 with different diagnostics but the results are not conclusive (Screen and Simmonds 2013; Barnes
110 2013; Riboldi et al. 2020), while the sinuosity of the midlatitude flow is projected to decrease in
111 future scenarios of CMIP5 models (Cattiaux et al. 2016).

112 Extreme temperature events occurrence can be influenced by change in the mean of temperature,
113 its variance or possibly by higher moments (Tamarin-Brodsky et al. 2019). Screen (2014) showed
114 that the subseasonal cold-season temperature variance has decreased in mid- to high-latitudes dur-
115 ing recent decades. The same is found in RCP4.5 scenarios of CMIP5 models by Gao et al. (2015)
116 which explain the large decrease in the number and duration of CS. In contrast, Cohen (2016)
117 found a slight increase of variance in midlatitudes from 1989 to 2015. Without changes in the at-
118 mospheric circulation, a decrease in the variance is expected simply because southward advected
119 Arctic air would be less cold (see e.g., Schneider et al. 2015). Therefore, to potentially get stronger
120 variance in a warming climate circulation changes are necessary.

121 The impact of AA on the mid-latitudes as suggested by Francis and Vavrus (2012) and Liu
122 et al. (2012) lacks strong theoretical arguments, as underlined by Wallace et al. (2014), Barnes
123 and Screen (2015) and Hoskins and Woollings (2015). The present study revisits this hypothesis
124 by analyzing in a simple modelling framework the effect of a decreased temperature gradient on
125 Rossby waves. We use the intermediate complexity model of Marshall and Molteni (1993): a

126 global three-level quasi-geostrophic model that has been shown to realistically simulate the win-
127 tertime Northern Hemisphere atmospheric circulation, its low-frequency variability (Corti et al.
128 1997; D’Andrea and Vautard 2001) and its synoptic-scale variability (Drouard et al. 2013; Rivière
129 and Drouard 2015).

130 Other idealized numerical studies investigated the impact of a decreased temperature gradient
131 on midlatitudes. Using a dry general circulation model, Hassanzadeh et al. (2014) found that a
132 reduced equator-to-pole gradient reduces the wind speed but also blockings and the meridional
133 amplitude of the waves. Based on the results of a hierarchy of models, Ronalds et al. (2018) found
134 that the narrowing of the jet on the poleward flank of its mean position in response to AA leads to a
135 decreased variability of the jet position. Using a zonally homogeneous forcing of the Marshall and
136 Molteni (1993)’s model, Robert et al. (2019) showed that a decrease in lower-level baroclinicity
137 makes the latitudinal fluctuations of the jet, more persistent.

138 In the present study, the same model is used but in a more realistic framework with topography,
139 land-sea contrasts and other zonally inhomogeneous forcings. An AA with an amplitude possible
140 for the end of the 21st century is included in the model’s forcing. Another difference with the
141 previously mentioned studies is that we focus on the changes in Rossby wave propagation and
142 magnitude and their impact on CS.

143 The basic equations of the model are introduced in Section 2, which also includes the setup of
144 two long-term numerical simulations: one *control* run with a climatology close to the present-
145 day wintertime observations and another forced by Arctic warming. The climatologies of the two
146 runs are analysed in section 2. In Section 3, diagnostics of Rossby wave spectral properties and
147 their differences between the two runs are described. Section 4 is dedicated to changes in CS. In
148 section 5 two additional integrations are analysed, in order to look at the sensitivity of the response
149 to the latitude of the forcing. Concluding remarks and discussions are provided in section 6.

150 2. Quasi-Geostrophic Model

151 a. Model equations

152 In this work, a version of the global quasi-geostrophic (QG) model first introduced by Marshall
 153 and Molteni (1993) is used. The model is particularly relevant here as it is the simplest baroclinic
 154 model to represent dry Rossby wave dynamics. It integrates the QG potential vorticity (q) equation
 155 at 3 vertical levels:

$$\frac{\partial q_i}{\partial t} = -J(\Psi_i, q_i) - D_i(\Psi_1, \Psi_2, \Psi_3) + S_i, \quad (1)$$

156 where the index i indicates the level (1: 200 hPa, 2: 500 hPa, 3: 800 hPa) and Ψ is the stream-
 157 function. J is the Jacobian operator and D_i aggregates all the dissipative terms (a linear drag at the
 158 lowest level, a vertical relaxation of temperature and a horizontal diffusion). The source term S_i –
 159 or forcing – is the subject of next section. At all times one can transform q into Ψ and vice-versa
 160 via a linear relationship. Discretized on the three levels, this reads:

$$\begin{aligned} q_1 &= 2\Omega \sin \varphi + \nabla^2 \Psi_1 - \frac{\Psi_1 - \Psi_2}{R_{12}^2} \\ q_2 &= 2\Omega \sin \varphi + \nabla^2 \Psi_1 + \frac{\Psi_1 - \Psi_2}{R_{12}^2} - \frac{\Psi_2 - \Psi_3}{R_{23}^2} \\ q_3 &= 2\Omega \left(1 + \frac{h}{H}\right) \sin \varphi + \nabla^2 \Psi_3 + \frac{\Psi_2 - \Psi_3}{R_{23}^2}, \end{aligned} \quad (2)$$

161 where we have introduced the Coriolis parameter $2\Omega \sin \varphi$ (φ is the latitude) and the two Rossby
 162 deformation radii ($R_{12} = 700\text{km}$ and $R_{23} = 450\text{km}$) for the two intermediate layers defined by
 163 two adjacent levels. In the lower level, the contribution of orography to the potential vorticity is
 164 introduced via the orographic height $h = h(\lambda, \varphi)$, depending on latitude φ and longitude λ , and a
 165 height scale H .

166 From the QG potential vorticity, all other variables can be diagnostically computed. The hori-
167 zontal wind is obtained from the streamfunction via geostrophy. The temperature is proportional
168 to the thickness of the layers (see appendix for its computation), and is thus defined at 650 hPa
169 and 350 hPa. The model is integrated in the spectral space, after discretization with a triangular
170 truncation of T42.

171 Since the model temperature is defined at 650 and 350 hPa, the fine scale structure of AA near
172 the surface is not represented. Our aim being to reproduce the effects of AA on the large-scale
173 meridional temperature gradient at midlatitudes, the use of the present model is however justified.

174 *b. Forcing*

175 The forcing at each of the three levels $S_i = S_i(\lambda, \varphi)$ is a space dependent and time constant
176 potential vorticity source, meant to represent the average effect of all the physical processes ne-
177 glected in the equations. These include diabatic heat sources, linked for example to precipitations,
178 cloud formation, the interaction with the ocean, etc., and the effect of the divergent flow.

179 In the original paper of Marshall and Molteni (1993), S_i was determined assuming that the PV
180 tendencies (computed as in Eq.(1)) starting from a large number of observed states have mean
181 equal to zero, which is equivalent to assuming that the large number of observed states is a stable
182 climatology of the model. This procedure can also be seen as an average correction of the model
183 error (D’Andrea and Vautard 2000).

184 In this work we use the methodology developed by Fromang and Rivière (2020), who applied
185 a variational algorithm to find an optimal S that minimises the difference between the model cli-
186 matology and a target one. Here, we used two target climatologies in order to define two different
187 forcing terms.

188 The control targeted climatology is defined as the wintertime (DJF) ERA-Interim (Dee et al.
 189 2011) streamfunction from 1979 to 2019. The Arctic Amplification (AA) targeted climatology
 190 is defined using the 6 winters (see Fig. 1) with the highest 850 hPa temperature in the polar
 191 cap regions ($\varphi > 70^\circ\text{N}$). Among the 6 selected winters, three of them feature a warming in the
 192 Barents-Kara sea region (2004-2005, 2005-2006 and 2011-2012), and three others display warm-
 193 ings centered in other regions: over the Greenland Sea region (2013-2014), between the Kara sea
 194 and the nearby continental Siberia (2015-2016), over the Bering Strait (2017-2018) (not shown).
 195 None of these winters present a pan-Arctic warming. The winter 2009-2010 is not selected even
 196 though it is well known for its warming in the south-west of Greenland (Overland et al. 2011; Cat-
 197 tiaux et al. 2010). Indeed, such a warming mainly occurs at latitudes outside the polar cap region
 198 here defined by $\varphi > 70^\circ\text{N}$. The streamfunction anomalies of the 6 chosen winters are averaged,
 199 multiplied by 3, and added back to the control targeted climatology. The multiplication by 3 is
 200 done to amplify the AA signal. Starting from these two targeted climatologies, two source terms
 201 S_{ctrl} and S_{AA} can be defined and are used to force two long-term (100 years of perpetual winter,
 202 so 36500 days) integrations of the model. The *ctrl* corresponds to the integration with the forcing
 203 S_{ctrl} , the AA run to the integration with the forcing S_{AA} . The obtained model climatologies are
 204 hereafter called *ctrl* and AA, respectively.

205 It is noteworthy that the difference of the two forcing terms (not shown) is not limited to the
 206 polar region, even if it is higher in this area. In fact, the variational methods does account for
 207 remote forcing effects that optimise the target climatology.

208 *c. Model Climatology*

209 The climatology and variability of the model, when forced with the S_{ctrl} and S_{AA} terms defined
 210 above are shown and described in the present subsection.

Figure 2 shows *ctrl* climatologies (black contours) and differences between AA and *ctrl* (shad-
ings) for the targeted climatologies (left column) and simulations (right column). The targeted
and model climatologies have great similarities by construction. Anomalous temperature patterns
(Figs. 2a,b) are obviously dominated by the AA, with a high peak centered on the Siberian side of
the North Pole (near 80°N, 60°E). Maps of zonal wind at 500hPa (Figs. 2c,d) show a huge circular
decrease around this peak. At lower latitudes, cold and hot anomalies are found, and temperature
anomaly, averaged zonally and over the 30°N-60°N band, is 0.48K for the difference between
the model climatologies, which matches well with the same value for the difference between the
targeted climatologies : 0.40K.

While the main features of these maps show remarkable correspondences, some differences re-
main. In the Asian - Western Pacific area, a large region of negative temperature anomaly exists in
ERA (Fig. 2a) which is also negative in the model but with much smaller amplitude (Fig. 2b). This
negative patch over the Eurasian continent has been shown to be related to the sea ice decrease
as recalled in the introduction (Honda et al. 2009; Mori et al. 2014). At the same longitudes but
further south, the anomalous temperature is positive which creates anomalous poleward-oriented
temperature gradient in Eastern Asia. In terms of zonal speed, the anomalous meridional temper-
ature gradient is associated with a northward shift of the western Pacific jet in ERA (Fig. 2c), but
with a more intense jet in the simulations (Fig. 2d).

In the Eastern Pacific area, the ERA and model climatologies are more similar with a temperature
increase mostly everywhere (Figs. 2a,b) leading to a poleward shift of the Pacific jet both in ERA
(Fig. 2c) and in the simulations (Fig. 2d).

In the area spanning the American to the European continent, ERA and the simulations show
more differences in the temperature maps (Figs. 2a,b). The southern positive temperature spot in
the Atlantic is shifted north and the American negative temperature spot is much weaker in the

simulation. For the zonal wind (Figs. 2c,d), the dipole of Fig. 2c, leading to a northern shift of the Atlantic jet in ERA is replaced by a decrease in the jet intensity, mainly on its southern flank, in the model. This quite large discrepancy between the targeted anomalies and the model anomalies is due to the iterative process being less successful in converging to the target AA climatology in that region. In the Pacific region, the differences between the targeted and model climatologies are of the same order of magnitude in the *AA* and *ctrl* cases.

One unexpected result from the selection of the 6 most extreme AA seasons is that it clearly shows a poleward shift of both the Pacific and Atlantic jets in ERA. This is surprising because many studies have reported equatorward shifts of the jets and a negative AO phase in response to AA (Cohen et al. 2010; Bader et al. 2011; Barnes and Screen 2015; Yang et al. 2016). As mentioned before, the well-known 2009-2010 winter marked by negative AO is not selected because its warming occurs at lower latitude than 70°N. By contrast, our selection picked up the 2013-2014 season which was more characterized by a positive AO (Rivière and Drouard 2015).

In summary, the jets are shifted poleward in ERA, while their intensity is changed in the simulations. In both, however, the zonal wind is reduced on the northern flank of the Pacific and Atlantic jets. The wind anomalies in the model are smaller than the ones in ERA: the decrease in zonal wind around polar regions is -11.8 m s^{-1} in ERA vs -6.3 m s^{-1} in the model, while the maximum value of Arctic amplification is comparable: 6.92 K in ERA against 6.85 K in the model. The changes in U and T in the simulations are 99% significant wherever the anomalies are non-zero using Welch's t-test.

In Fig. 3, the standard deviation of the low- and high-frequency meridional wind at 500 hPa are shown. The filter is defined by a seven-point Welch window applied to daily datasets, which has a 10-day cutoff period. Again, the contours correspond to the *ctrl* case while color shadings report the AA response. The standard deviation changes are 99% significant using a Levene test.

259 The standard deviations of the low- and high-frequency meridional wind in the *ctrl* simulation
260 can be compared to those of ERA-I reanalysis (see e.g. Fromang and Rivière (2020) for more
261 details). Two peaks in the standard deviation of the low-frequency meridional wind are present in
262 the *ctrl* simulation (Fig. 3a), one located in the Northeastern Pacific south of Alaska and another
263 one in the Northeastern Atlantic north of the British Isles. These peaks, found downstream of
264 the two main storm-track regions shown in Fig. 3b, are roughly located at the same location as in
265 the reanalysis but their amplitudes are weaker (almost one third reduction compared to ERA-I).
266 Likewise, the high-frequency variability associated with the storm tracks is also less intense than
267 in ERA, especially in the North Atlantic sector.

268 The difference in low-frequency meridional wind standard deviation between *AA* and *ctrl* sim-
269 ulations is generally positive with an enhancement of 0.714 m s^{-1} (12%) in the 30°N - 60°N band
270 (shadings in Fig. 3a). The positive response shows three peaks, two over North America and one in
271 the central North Atlantic. The peak of standard deviation in the Northeastern Pacific is stronger
272 in the *AA* run by 20 %, but the peak in the Northeastern Atlantic is shifted westward without much
273 change in its amplitude. The increase in low-frequency variability is not systematic. There is a
274 region with negative anomaly forming a crescent from Scandinavia to a large part of Northern
275 Eurasia in the same area where the strongest decrease in zonal wind is observed (Fig. 2d).

276 The anomalous *AA* composite of the high-frequency standard deviation (Fig. 3b) is mainly re-
277 lated to the midlatitude jets changes (Fig. 2d). The Pacific storm track intensifies in its western
278 part and then increases on its poleward flank further east in a similar manner to the westerlies.
279 The Atlantic storm track generally decreases in intensity. This is consistent with a decrease in
280 the intensity of the upper-level Atlantic jet because it is accompanied by a decrease in baroclin-
281 icity. Generally speaking, the negative values cover larger areas than the positive ones. There
282 is a well-defined large-scale region of negative values from Scandinavia to Northern Eurasia as

for the low-frequency variability. However, there are also decreases in high-frequency eddy activity in regions where the low-frequency variability increases: this is the case of the latitudinal band 30°N–40°N extending from 140°W to 0°E. In this band the variability shifts from high- to low-frequency suggesting lower-frequency waves and slower phase speed.

A systematic feature therefore emerges at midlatitudes from Figs. 3a,b: the low-frequency contribution to the total variability increases. This is confirmed in Fig. 3c. The total variance, denoted as σ_{tot}^2 , is nearly the sum of the variance of the low- and high-frequency meridional wind, denoted as σ_{LF}^2 and σ_{HF}^2 respectively, as the cross-covariance between the low- and high-frequency components is small. Hence, $\sigma_{tot}^2 \approx \sigma_{LF}^2 + \sigma_{HF}^2$. Figure 3c presents the percentage of variance explained by the low-frequency component: $100\sigma_{LF}^2 / \sigma_{tot}^2$. For the *ctrl* run (contours), the percentage fluctuates between 20% and 50% at midlatitudes and these regions are logically dominated by storm-track and high-frequency activity. In the polar region, the reverse happens and the transient eddy activity is dominated by the low-frequency component.

Compared to the *ctrl* case, the *AA* run shows a systematic increase of the low-frequency contribution to the total variance everywhere at midlatitudes. Even in the Eurasian area, where the low-frequency variance decreases, the high-frequency decreases more, so the impact of the low-frequency is proportionally higher. In the Arctic, where the part of the low-frequency is the highest in the *ctrl*, the low-frequency contribution slightly decreases in the *AA* run.

Another result emerges from Figs. 3a,b: as first noted by Hassanzadeh et al. (2014), the total variance varies in tandem with the zonal wind. The high- and low-frequency variances (and thus the total variance) increase where the zonal wind strengthens (South of Alaska) and decrease where the zonal wind weakens (Eurasia). The spatial correspondence with the zonal wind anomalies tends to be more obvious for the high-frequency variance, which probably comes from the close relationship between baroclinicity and storm-track activity as emphasized above.

307 To conclude, the most original result brought out by this section is an increased percentage of
308 variance explained by the low-frequency component at the expense of the high-frequency compo-
309 nent at midlatitudes. That result is further detailed and interpreted in the following section.

310 **3. Rossby wave response**

311 *a. Space-time spectral analysis*

312 Hayashi (1971) has introduced a spectral analysis of space-time variability for geophysical vari-
313 ables on a latitude circle. Rather than a usual spectrum with the frequency ω in abscissa and the
314 dimensionless zonal wavenumber m in ordinate, in this spectrum the direction of propagation is
315 included. Particularly, in Hayashi (1979), ω is replaced by the phase speed c_ϕ , which is negative
316 in the case of westward waves. Hayashi (1981) proposed to separate the field between transient
317 and standing (stationary) waves, the last being the sum of westward and eastward waves having
318 the same speed and phase.

319 The power density spectrum is operationally computed as presented by Dell'Aquila et al. (2005)
320 (see their appendix) but using the meridional wind rather than the geopotential height. The first
321 step is the Fourier analysis of the spatial field, and the second is a time power spectrum of each
322 wavenumber. Then the dependency to phase speed is found using $c_\phi = \frac{\omega}{k} = \frac{\omega a \cos \varphi}{m}$, where k
323 is the dimensional zonal wavenumber, m the dimensionless zonal wavenumber and a the Earth
324 radius. Here, Hayashi spectra are computed with meridional wind at 500 hPa and finally averaged
325 over the latitudinal band 35°N-65°N.

326 An alternative method to visualize the spectral characteristics of the signal is the phase speed-
327 amplitude histogram. Here the model is run for 10 years with a sampling frequency of 6 hours
328 instead of 1 day to get accurate computation of the phase speed. By applying a Fast Fourier

329 Transform (FFT) to the latitude-averaged meridional wind, we get the amplitude and the phase ϕ
 330 for each wavenumber and time. The phase speed is computed as follows:

$$c_\phi = \frac{\omega}{k} = \frac{a \cos \phi \Delta \phi}{m \Delta t}, \quad (3)$$

331 where Δ represents a centered scheme difference with 6-hourly outputs.

332 *b. Results*

333 Figure 4 shows the Hayashi spectrum of the *ctrl* simulation (contour), and the difference of
 334 spectrum between the *AA* and *ctrl* simulations (color). For synoptic wavenumber range (5-9) the
 335 *ctrl* spectrum shows high power for eastward travelling waves. At smaller wavenumbers (less than
 336 4), the power peaks near the zero phase velocity. For wavenumbers 3 and 4, the power is still
 337 stronger for eastward travelling waves than westward travelling waves while for wavenumbers 1
 338 and 2 westward travelling waves have power as high as the eastward travelling waves. In general,
 339 these features compare well with observed spectrum (see Dell’Aquila et al. 2005, and references
 340 therein) and testify that the QG simulations have realistic Rossby wave dynamics.

341 The difference between the *AA* and *ctrl* simulations shows an increase of power for negative and
 342 zero phase speeds which is particularly strong for wavenumbers 3 to 5, and a small decrease in
 343 power for positive phase speeds. In particular, the power of quasi-stationary waves for wavenum-
 344 bers 4 and 5 has increased by nearly 50%, while that of modes with wavenumbers 2 and 3 has been
 345 reduced. This shift toward higher wavenumbers is expected from Hoskins and Woollings (2015)’s
 346 argument based on the stationary Rossby wave barotropic formulation and its dependence on zonal
 347 wind.

348 Phase speed - amplitude histograms are shown in Figure 5. Histograms of the *ctrl* simulation are
349 in agreement with the Hayashi spectrum of Fig. 4: waves travel more and more rapidly eastward
350 for higher wavenumber. For wavenumbers 1 to 3 the peak of the histogram is close to zero, i.e.
351 the waves are mostly stationary. For higher wavenumber the peak moves more and more to the
352 positive phase speed side (between 0 and 5 m s⁻¹ for m=3-5 and 5-10 m s⁻¹ for m=6-8).

353 Differences between *AA* and *ctrl* runs exhibit dipolar anomaly patterns for all wavenumbers.
354 There are more cases with near-zero or even negative phase speed and less cases with positive
355 phase speed in presence of *AA* for all wavenumbers except for $m = 1$ which is not very ener-
356 getic. Again in agreement with the Hayashi spectrum, wavenumbers 3 to 5 gain power through
357 an increase in the number of high-energy days with near-zero phase speed, while decreases in
358 frequency are for eastward or low amplitude waves. The negative and positive anomalies for east-
359 ward and westward phase speed respectively are more symmetric in the histograms than in the
360 Hayashi spectrum. The latter is indeed more dominated by the increase in power of near-zero and
361 negative phase speed modes. This can be understood by the fact that the increased number of days
362 generally corresponds to high-amplitude waves with high power while the decrease corresponds
363 to low-amplitude waves whose impact on the power is much smaller.

364 In summary, the two diagnostics give coherent results, showing an increase of westward and
365 quasi-stationary waves in number and in power, and a decrease of eastward waves, particularly for
366 long waves with wavenumbers 3 to 5.

367 *c. Interpretation using the barotropic Rossby wave dispersion relation*

368 To understand the decrease in phase speed at midlatitude, we use the barotropic dispersion re-
369 lation of a linear Rossby wave evolving along a zonal background flow slowly varying in the
370 meridional direction y :

$$c_\phi = U - \frac{\beta - \frac{\partial^2 U}{\partial y^2}}{k^2 + l^2}, \quad (4)$$

where k and l are zonal and meridional wavenumbers, and U the background zonal wind. Such a relation has been extensively used to study the meridional propagation of Rossby waves in general (Hoskins and Karoly 1981; Held 1983; Hoskins and Ambrizzi 1993) and more recently in the context of the impact of polar amplification on midlatitudes (Ronalds et al. 2018). Here, our purpose is to use this relation to attribute changes in the phase speed for a given k to specific terms of the equation. For a given k , the difference in phase speed between the *AA* and *ctrl* runs Δc_ϕ can be expressed at first order as:

$$\Delta c_\phi = \Delta U - \frac{\Delta \left(\beta - \frac{\partial^2 U}{\partial y^2} \right)}{k^2 + l^2} + \frac{\beta - \frac{\partial^2 U}{\partial y^2}}{k^2} \frac{\Delta \left(\frac{l^2}{k^2} \right)}{\left(1 + \frac{l^2}{k^2} \right)^2}. \quad (5)$$

The first term on the right hand side of the equation corresponds to changes in the mean wind, the second term to changes in the absolute vorticity gradient, and the third term to changes in the meridional wavenumber l .

The first and second terms of the right hand side of Eq. (5) can be readily computed from changes in the climatological means of the zonal wind and absolute vorticity gradient (Figs. 6a,b). The background zonal wind U is thus averaged over time and longitude but also vertically averaged over the three levels of the model. The ratio l/k is estimated by computing the ratio of the standard deviation of the zonal and meridional geostrophic wind components: $l/k = \sqrt{\sum (u_k - \bar{u}_k)^2} / \sqrt{\sum (v_k - \bar{v}_k)^2}$ where u_k and v_k denote the zonal and meridional geostrophic wind components of a given k , respectively. The bars correspond to the time mean. All the terms of Eq. (4) averaged over boxes of 15° in latitude are shown in Figs. 6c,d.

The increase in temperature in the polar cap region decreases the zonal wind on the northern flank of the jet, north of 55°N (Figs. 6a,b). This decrease occurs in both ERA (with the anomalies

multiplied by 3) and simulations. At lower latitude, ERA and simulations show more differences, consistently with our previous comments on Fig. 2: in ERA (Fig. 6a), the jet is shifted northward with AA whereas not much difference is detected in the simulations, apart a very slight decrease in amplitude. In both ERA and simulations, however, the zonal wind decrease on the poleward flank of the jet makes the jet thinner.

The different terms of Eq. (5) are shown in Figs. 6c,d, computed for zonal wavenumber $m = 5$ and for the different latitudinal bands. The first term of Eq. (5), associated with zonal wind decrease, appears to be the main driver of the phase speed variations at high latitudes (see the blue bars for latitudes greater than 50°N). The second term of the equation is linked to changes in the absolute vorticity gradient, or the second derivative (i.e the convexity) of the zonal wind. For lower latitudes (in the band from 35°N to 50°N), as zonal wind decreases on the northern flank of the jet, the jet becomes thinner, so the absolute vorticity gradient increases in the jet-core region. This makes the second term of Eq. (5) negative and has the effect of reducing the phase speed. In ERA, the second term completely offsets the increase of zonal wind in the latitudinal band 35°N – 50°N , and consequently the phase speed is reduced. In the simulations, the second term is also negative, although less strongly than in ERA, but its net effect, in conjunction with the third term, is also a decrease in phase speed. The third term of the equation is linked to changes in meridional wavenumber for a given zonal wavenumber. Smaller l values correspond to more meridionally elongated eddies. In the band 35°N – 50°N , the decrease of l contributes to the phase speed decrease for both reanalysis and simulations.

Finally, south of 35°N , the models and the reanalysis behave differently: there is no net change in the model while the reanalysis exhibits an increase in the phase speed according to the considered barotropic framework. It is important to emphasize that changes of the perturbation elongation Δl cannot be directly related to the Francis and Vavrus (2012) argument about the elongation of the

415 jet meanders. Indeed, Δl measures the change in the perturbation meridional extension where the
416 perturbation is defined as the deviation from the time mean, so here the effects of the modified time
417 mean flow on the transient perturbation is estimated. Francis and Vavrus (2012) considered the
418 time mean flow anomaly as a part of the perturbation to explain the increased meridional extension
419 of the jet meanders: they did not consider the potential change of the transient perturbation itself.

420 The above results are shown for wavenumber 5, but similar results are obtained for zonal
421 wavenumbers 3 to 8 (not shown): phase speeds are found to decrease at midlatitudes (35°N-65°N).
422 The comparative impact of the three terms discussed above, however, changes with wavenumber.
423 Higher wavenumbers are more strongly affected by zonal wind changes while lower wavenumbers
424 are more largely influenced by PV gradient or meridional elongation changes.

425 To conclude, in both reanalysis and simulations, the change in synoptic Rossby wave phase
426 speed at high latitudes is dominated by the change in zonal wind, while at lower latitudes, near the
427 jet core, it is more impacted by the variations of the absolute vorticity gradient or the meridional
428 elongation. At all latitudinal bands, the net effect is a decrease in phase speed according to the
429 barotropic dispersion relation. Even though the change in the eddy elongation is not explained,
430 the effect of AA on the background zonal wind and absolute vorticity gradient, explained by the
431 narrowing of the jet, provides an interpretation for the slowing down of Rossby waves.

432 **4. Cold Spells**

433 In this section, the impact of variability and wave activity changes in the model on extreme
434 phenomena is examined by focusing on cold spells (CS). A CS is here defined as an episode of at
435 least 5 consecutive days below a temperature threshold depending on the geographical position.
436 At each grid point the threshold is established as the 10th percentile of the temperature distribution
437 of the *ctrl* simulation.

438 In the *ctrl* run, there is a large number of CS in the polar region, with more than 200 CS in
439 total during the 100 years of the run (contours in Fig. 7a). Minima are found at midlatitudes in
440 the Pacific and Atlantic areas, with less than 40 CS. This is linked to the storm tracks, so that
441 temperature anomalies tend to be short lived, while CS duration is at least 5 days by definition.
442 The number of CS slightly increases over continents compared to the ocean basins to reach values
443 of around 100 CS over Northern America, Europe and Asia.

444 Anomalies in the *AA* run are dominated by a mean decrease of 60% in CS in the Arctic region
445 (60°N - 90°N). In midlatitudes, the frequency of CS increases in 3 areas: North America, Western
446 Europe and Central Asia. This map closely resembles the mean temperature anomalies spatial
447 distribution (Fig. 2b): a locally hotter or colder mean temperature obviously influences the prob-
448 ability of exceeding the threshold. This is reflected in the correlation coefficient between the two
449 maps, which is equal to -0.89.

450 In Fig. 7b, the mean duration of CS is computed as the number of days in Cold Spells divided
451 by the number of Cold Spells. CS always last more than 5 days by definition, but are shorter in
452 the storm tracks and last longer in the polar region. The changes in CS duration are related to
453 changes in CS number, with a large decrease at high latitude and an increase at midlatitudes (0.1
454 days in average). In some midlatitude regions, the increase is large as in the Atlantic sector where
455 the low-frequency variance is strong too (Fig. 3a).

456 We can separate the effect of the mean and the effect of the variability by subtracting the mean
457 temperature difference to the *AA* temperature data sets before applying CS definition, and by
458 applying the same standard deviation to both time series (for each grid point) respectively. This

459 leads to the definitions of two new temperatures

$$T_{=mean} = T_{AA}(\lambda, \varphi, t) + \bar{T}_{ctrl}(\lambda, \varphi) - \bar{T}_{AA}(\lambda, \varphi), \quad (6)$$

$$T_{=\sigma} = (T_{AA}(\lambda, \varphi, t) - \bar{T}_{AA}(\lambda, \varphi)) \frac{\sigma(T_{ctrl}(\lambda, \varphi, t))}{\sigma(T_{AA}(\lambda, \varphi, t))} + \bar{T}_{AA}(\lambda, \varphi), \quad (7)$$

460 where σ stands for the standard deviation. The number of CS defined as the number of episodes
 461 for which $T_{=\sigma}$ (resp. $T_{=mean}$) is below the 10th percentile of the temperature distribution of the *ctrl*
 462 simulation for at least 5 consecutive days is shown in Fig. 8a (resp. Fig. 8b). When the influence
 463 of the standard deviation on the definition of CS is removed (Fig. 8a), the impact of the changes
 464 in the mean temperature is highlighted by comparing Fig. 8a and Fig. 8c. The correspondence
 465 between Fig. 8a and Fig. 8c is compelling and it increases with respect to the full response shown
 466 in Fig. 7a. The correlation coefficient slightly increases to -0.91.

467 In Fig. 8b, the contributions of the higher moments of the temperature distribution to the CS can
 468 be assessed. When the differences in the mean are suppressed, smaller differences between the AA
 469 and *ctrl* simulations appear in the polar cap region, and the link with the low-frequency tempera-
 470 ture standard deviation (shown in Fig. 8d) appears : the correlation coefficient is 0.83. The three
 471 regions of particular increase in CS number in this map, that are the North Pacific, North America,
 472 North Atlantic, correspond to regions of important increase in the low-frequency temperature stan-
 473 dard deviation. Conversely, the crescent-shaped region of decrease in the number of CS between
 474 Scandinavia and Siberia corresponds to a region where the low-frequency temperature variabil-
 475 ity decreases (Fig. 8d), as has been already discussed in section 2c. Because of the condition on
 476 the minimal duration of CS (5 days), CS without the mean (Fig. 8b) fits better with changes in
 477 the low-frequency variability than with the total variability. Indeed, the correlation with the total
 478 standard deviation (not shown) is 0.76 and less than with the low-frequency component (0.83).
 479 This is particularly true in the band 30°N-40°N extending from the eastern Pacific to the eastern

480 Atlantic where there is no increase in the total variability but an increase in both the low-frequency
481 variability and the number of CS.

482 It appears clearly that the increased variability can locally counterbalance the mean temperature
483 increase and even offset it completely locally, particularly in the midlatitude band. For instance,
484 over continental North America the mean temperature difference of the two simulations is near
485 zero (Fig. 8c), but there are more CS (Fig. 7a) because the low-frequency variability increases
486 (Fig. 8d) and causes long events of polar air advection toward the US.

487 For illustrative purposes, we now focus on two particular regions in the NH: one in Western
488 Europe and one in the North American Midwest, marked by the two green crosses. Mnemonically
489 we can identify them with Brest, France, and with Winnipeg, Manitoba. For both locations, we
490 report in Table 1 the changes in the total number of CS and the changes following the definitions
491 of Eq. (7) and Eq. (6). These estimates are obtained as follows. If one takes the example of
492 Winnipeg, there are 138 CS in the *ctrl* simulation, and 278 CS in the AA simulation, hence an
493 increase of 140 CS is reported in Table 1 (compare also with Figure 7a). Without the influence of
494 the mean temperature change, i.e. using Eq. (6), we found 241 CS in the AA simulation (Figure 8a),
495 hence an increase of 103 CS compared to the *ctrl* run (fourth column in Table 1). To summarize,
496 the total change is 140 CS in the AA run, of which 103 CS are not due to the mean. Making the
497 hypothesis that all the others, i.e. $140 - 103 = 37$ CS, are due to the mean change, we attribute 26%
498 of the CS number increase to the mean temperature change. Applying the same line of reasoning
499 when using Eq. (7), we found that 60% of the changes of CS in Winnipeg are due to the variance
500 increase. We applied the same method for Brest, where we found that it is more the change in the
501 mean (50%) than the variability (17%) that matters most for the CS increase at that location.

502 Table 1 also shows the mean duration of CS. In Brest, the mean duration of CS increases by
503 0.2 days, In Winnipeg, it extends by 0.5 days. While in Brest the change of duration is due to the
504 mean decrease of temperature, in Winnipeg its impact appears to be minimal.

505 In summary, changes in mean temperature dominates the changes of CS, but the response to AA
506 can balance these modifications in some areas where the changes in low-frequency temperature
507 standard deviation are large, and where changes in the mean are consequently overwhelmed by
508 standard deviation increase. This is for example the case over North America. It is worth noting
509 that individual CS events last also, on average, longer at midlatitudes in presence of AA. Both the
510 increase in the number of CS (as defined with a minimum duration of 5 days) and the increased
511 duration of CS in presence of AA are attributed to an increase in low-frequency variability in close
512 connection with a slowing down of the eastward propagation of Rossby waves.

513 **5. Sensitivity to tropical and extratropical regions**

514 To assess the relative importance of the different regions in setting up the changes in Rossby
515 wave spectra and CS statistics in the AA run compared to the *ctrl* run, two additional 100-year-
516 long simulations were performed. To do so, two forcing S were calculated using the iterative
517 method described in section 2: for the first one, referred to as the *extratrop* simulation, the target
518 climatology corresponds to the control target climatology south of a latitude of 35°N, but to the
519 AA target climatology for latitudes larger than 35°N. The choice of this particular latitude is made
520 following Fig.6 which shows that the latitudinal band 35°N-50°N matters in our interpretation of
521 the phase speed decrease. For the second simulation, referred to as the *trop* simulation, the target
522 climatology is set to the control target climatology, except for latitudes between 35°S and 35°N
523 where it corresponds to the AA targeted climatology. If we discard the influence of the Southern

524 Hemisphere south of 35°S , and expect a linear response for the two runs, their sum should provide
525 the AA run.

526 The temperature anomalies of the two simulations with respect to the *ctrl* run are shown in
527 Fig. 9a,b. As expected, the *extratrop* simulation exhibits a warming pattern at high latitudes centred
528 between the North Pole and Siberia of the same amplitude as in the AA simulation (compare Fig. 9a
529 with Fig. 2b). In mid latitudes, the *extratrop* simulation patterns display less similarities with the
530 AA run, which might be due to the closeness with the transitional latitude 35°N . As expected from
531 the minimization algorithm, the *trop* simulation exhibits less strong anomalies than the *extratrop*
532 simulation at high latitudes. However, the anomalies are not small and there is a warming pattern
533 centred over the pole (Fig. 9b) whose peak amplitude is half that of the AA or *extratrop* runs. It
534 shows that, despite the minimization procedure which tends to avoid any anomalies with respect
535 to the *ctrl* simulation at high latitudes, the model forms anomalies with a clear Arctic warming
536 pattern from the sole knowledge of tropical anomalies. It supports the recent findings of Baggett
537 and Lee (2017) and Gong et al. (2019) for instance who showed that planetary-scale Rossby wave
538 trains triggered by anomalous tropical convection in the western Pacific are responsible for setting
539 the Arctic warming patterns.

540 Hayashi spectra for the two sensitivity experiments are shown in Fig. 10. The spectrum of the
541 *extratrop* simulation exhibits a dipolar anomaly corresponding to a shift towards smaller – or more
542 negative – phase velocities, which is more obvious for wavenumbers 3 and 4 (Fig. 10a). This shift
543 toward lower frequencies corroborates the interpretation provided in section 3c. The spectrum of
544 the *trop* simulation is similar to the AA run with increased energy for near-zero and westward phase
545 speed (compare Fig. 10b and Fig. 4). It means that the mean tropical anomalies are responsible for
546 an increased activity of the slowly and slightly westward propagating waves at mid latitudes. An
547 interpretation for such a behavior would need further analysis and is not provided here. Finally,

the two latitudinal bands participate in the energy spectra of the *AA* run but the *trop* simulation brings more similarities with the latter run.

CS statistics for the two simulations are shown in Fig.9. As for the *AA* run, the change in CS number is dominated by the change in the mean temperature (Figs. 9c,d). By subtracting the mean temperature difference as in Eq. (6), the CS number largely increases in midlatitudes in the *trop* simulation as in the *AA* run but more importantly in the Pacific than in the Atlantic while the reverse happens for *AA* run (see Figs. 9f and 8b). For the *extratrop* simulation, there is some increase in CS number in a narrow latitudinal band between 30°N and 50°N but further north the decrease is rather systematic in all the regions.

To conclude, it is the *trop* simulation rather than the *extratrop* simulation that brings more similarities with the *AA* run in terms of change in CS number and change in energy spectra. The net increase in near-zero and small westward phase speed in the *trop* simulation provides an explanation for the increased CS number in mid latitudes when the mean temperature is the same as in *ctrl*. Yet the reason for such an increased amplitude for slow westward phase speed remains an open question that would need further analysis. The *extratrop* simulation exhibits a more moderate increase in CS number in a narrower latitudinal band that can be explained by the shift from high to low frequencies resulting from the background flow changes according to section 3c. But it seems of secondary importance relative to the signal coming from the tropics.

6. Discussion and Conclusion

In this article, a dynamical mechanism linking Arctic Amplification and Rossby wave dynamics has been studied using a simple, quasi-geostrophic model. The link to the frequency and intensity of Cold Spells has also been addressed. While a comprehensive theory on the wave response to the northern midlatitude to AA is still missing (Hoskins and Woollings 2015), advances have

recently been obtained by the use of such highly idealised models. For instance, the reduction of the hemispheric meridional temperature gradient has been shown to reduce the frequency of blocks (Hassanzadeh et al. 2014) and to increase the persistence of the leading mode of variability of midlatitude jets (Robert et al. 2019). We meant here to bring the level of model complexity one step further and introduce longitudinal details (continents, orography), and account for the fact that AA does not really determine a latitudinally uniform temperature gradient reduction.

Our strategy was to compare two runs, one serving as a reference and another one where anomalies corresponding to warm Arctic winters are imposed everywhere over the globe. Differences in terms of Rossby waves properties and cold spells are then analyzed. The imposed anomalies cause a reduction of planetary and synoptic Rossby wave phase speed at midlatitudes. Such changes are in agreement with changes in the variances of temperature and wind for different frequency bands: high-frequency variability (periods smaller than a week) mainly decreases while low-frequency variability (periods larger than a week) increases.

The slowing down of Rossby wave phase speed can be interpreted using the barotropic Rossby wave dispersion relation and background flow anomalies at mid and high latitudes. The polar increase of temperature above 60°N weakens the zonal wind at those latitudes, which correspond on average to the northern flank of the jet. This makes the jet thinner, which changes the meridional PV gradient as a result. Both these effects decrease the phase speed of Rossby waves, the first being dominant above 60°N and the second just south of it, closer to the jet core. Such an interpretation is confirmed when only anomalies north of 35°N are imposed in the model. The combined effect of the zonal wind and PV gradient anomalies provide a refinement of the Francis and Vavrus argument.

To be more precise, the increase in amplitude of slowly propagating waves is larger relative to the decrease of rapidly eastward propagating waves. This effect appears to originate from the tropical

anomalies present during the warm Arctic winters. When the model is forced to converge toward tropical anomalies only, it does indeed create a warm Arctic pattern – which confirms the findings by Baggett and Lee (2017) and Gong et al. (2019) – but additionally it shows a large increase in slowly-propagating wave energy similar to the full AA run. The reason for that behavior remains unclear.

We have also analyzed the impact of AA on Cold Spells (CS). CS are defined with a temperature threshold-based criterion, complemented by a duration criterion of at least 5 days. The increase of mean temperature has the obvious effect of decreasing on average the probability of passing the threshold, and hence decreases the average occurrence of CS. Nevertheless, the local increase of low-frequency variability associated with the change of Rossby wave dynamics opposes in some regions the effect of the mean temperature increase. There is a compensation of the two effects in some mid-latitude regions as in the northwestern Atlantic south of Greenland, which leaves the frequency of CS little changed. A net increase in CS frequency appears in regions of colder mean temperature. The increase over Western Europe is largely due to the colder mean temperature but also slightly due to increase in the low-frequency variability. The increase in CS in the North American continental regions is more largely due to the increase in variability than to the colder mean temperature. Additionally, the increase in low-frequency variability explains why CS with duration longer than 5 days are more numerous. Numerical experiments comparing the response to different latitudinal bands of anomalies show that the increased number of CS is more due to the tropics than the mid- and high-latitudes. This is consistent with the fact that the amplification of slowly-varying waves is more important in the run where tropical anomalies are imposed.

Confirming the above results by looking at the observations of the real atmosphere is not an easy task. First, the real anomalies of the reanalysis data are small compared to those used to force the model. Remember that the warm winter forcing in the model is 2 to 3 times greater than that of the

619 real data. Second, the observational period covers a much smaller period than that analyzed in the
620 model and cannot provide robust statistical results. Third, our analysis is based on the composite of
621 6 very different seasons and the net effect might be small compared to individual seasons. Indeed,
622 the recent review article of Cohen et al. (2020) shows that the observed large scale anomalies
623 linked to regionally localized Arctic anomalous warmings can be very different from one region
624 to another. For example the large scale response of the Asian continental temperature to a heating
625 of the Kara sea or of the Greenland sea can be opposed in sign. The response shown in the present
626 work could consequently be real but not be the prevailing one.

627 Exploring each of these localized forcings and their causal relationship with different heating
628 patterns in the tropics (which we found in the present paper to be important) can definitely be
629 done in the present modelling framework. This would require a a full study by itself which we
630 reserve for future work. This would also help clarify whether global warming and other regional
631 warmings, and not only anomalies setting the polar amplification pattern, could also induce a net
632 change in Rossby wave phase speed.

633 APPENDIX

634 Temperature computation in QG model

635 The only free variable of the QG model is the QG potential vorticity q (or equivalently the
636 streamfunction Ψ). Temperature can be diagnostically computed from it. Using Ψ and the Coriolis
637 parameter f , the geopotential Φ satisfies the balance equation (see e.g. Holton (2004) or any other
638 geophysical fluid dynamics manual)

$$\nabla^2 \Phi = \nabla \cdot (f \nabla \Psi) = \nabla f \cdot \nabla \Psi + f \nabla^2 \Psi \quad (\text{A1})$$

639 The Laplacian of the geopotential can then be integrated twice to get the geopotential. The
 640 computation of the geopotential consequently involves an arbitrary integration constant that cor-
 641 responds to its spatial mean value.

642 The temperature between the levels is computed using two assumptions: the hydrostatic approx-
 643 imation and the law of perfect gas for air, which give:

$$\delta\Phi = -R_{air}T\frac{\delta P}{P} \quad (A2)$$

644 And discretizing on the vertical axis of the model:

$$T_{12} = \frac{\Phi_1 - \Phi_2}{R_{air} \ln(\frac{P_2}{P_1})} \quad (A3)$$

$$T_{23} = \frac{\Phi_2 - \Phi_3}{R_{air} \ln(\frac{P_3}{P_2})}. \quad (A4)$$

645 T_{12} is the mean temperature between levels 1 (200 hPa) and 2 (500 hPa), that is at the mean
 646 pressure level 350 hPa. T_{23} is the mean temperature between levels 2 (500 hPa) and 3 (800 hPa),
 647 that is at the lower level 650 hPa. R_{air} is the gas constant for dry air.

648 The temperature computed using this method depends on the arbitrary integration coefficient of
 649 Φ from the inversion of the horizontal Laplacian in Eq.(A1). To define this constant we proceed
 650 as follows: for the *ctrl* integration we chose the mean geopotential height of the US standard
 651 atmosphere; for the *AA* integration we chose an *ad hoc* constant, so that the space-mean difference
 652 of temperature between the two integrations is the same of the two target climatologies (i.e. the
 653 map in color in Fig. 2a, and the equivalent map at 350 hPa). The maps in Figs. 2a and 2b,
 654 consequently, have the same global space mean by construction.

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885 mean in the same (Fig. 8b, colors), and when the standard deviation in the same (Fig. 8a, colors), percent of
886 the change due to the mean and the standard deviation, and changes in mean temperature (Fig. 8c, colors) and
887 temperature standard deviation (Fig. 8d, colors). The last columns compare CS durations for *ctrl* and AA runs
888 (Fig. 7b) and when the mean is the same.

Position	CS number	Change of CS			Due to		Anomaly (K)		CS Duration (days)		
	<i>ctrl</i> run	total	same mean	same std	mean	std	mean	std	<i>ctrl</i>	AA run	same mean
Brest	79	+159	+80	+132	50%	17%	-1.01	+0.25	6.44	6.64	6.43
Winnipeg	138	+140	+103	+56	26%	60%	-0.34	+0.64	5.99	6.53	6.49

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933 **Fig. 10.** Same as Fig. 4 but for (a) *extratrop-ctrl* difference and (b) *trop-ctrl* difference. 55

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a) T 650 hPa *extratrop-ctrl*

b) T 650 hPa *trop-ctrl*

c) CS number *extratrop-ctrl*

d) CS number *trop-ctrl*

e) CS same mean of T *extratrop-ctrl*

f) CS same mean of T *trop-ctrl*

FIG. 9. (Upper panels) 650-hPa temperature anomaly, i.e same as Fig. 2b but for (a) *extratrop-ctrl* difference and (b) *trop-ctrl* difference. (Middle panels) CS number, i.e same as Fig. 7a but for (c) *extratrop-ctrl* difference and (d) *trop-ctrl* difference. (Bottom panels) CS number with data translated to the same mean as in *ctrl*, i.e same as Fig. 8b but for (c) *extratrop-ctrl* difference and (d) *trop-ctrl* difference.

FIG. 10. Same as Fig. 4 but for (a) *extratrop-ctrl* difference and (b) *trop-ctrl* difference.