

We thank the three referees for carefully reading our manuscript, and for their constructive comments. In the following we will respond to the various comments and point out any changes we made to the paper based on them. Line numbers and figure references in the reviewer's comments refer to the original manuscript, while all references in the responses refer to the revised version of the manuscript with tracked changes (from the originally submitted version, shown in red/blue), which can be found at the end of this document. The comments of the referees are in black italics; our responses are in blue.

## **Response to Referee 1:**

*The authors report on a set of baroclinic lifecycle experiments aimed at better understanding the influence of the stratospheric vortex on tropospheric jets. Consistent with previous works, they find that life cycles are substantially modified by presence or lack of a stratospheric vortex. They point out the importance of the winds in the lower stratosphere and identify an amplifying effect of surface friction on the near surface response.*

### *Discussion paper*

*The experiments are interesting and potentially quite informative - more idealized experiments such as these are, in my opinion, under-utilized for studying this difficult problem. In particular I found them to be a very clear illustration of the importance of non-linear phase of baroclinic lifecycles in understanding the nature of the tropospheric eddy feedbacks. However, I found some of the discussion to be a bit unsatisfying. I think the most useful suggestion I can give is to strengthen and clarify the comparison of the present results with previous studies on baroclinic lifecycles. Section 1.2 does set up some past results, but it's never made that clear how the present results fit in.*

*For instance: the Wittman papers emphasize changes in linear growth rates, while the present paper emphasizes the non-linear decay phase as the key period of difference, finding (from what I can tell) relatively weak changes in the growth rates. Is this discrepancy consistent with the Smy and Scott emphasis on PV in the sub-vortex region? How much does this have to do with the different basic states? To be clear - overall this is a solid piece of science and should ultimately be worthy of publication. I have some further specific comments and suggestions below; some of which might involve significant further efforts. I do not require all of these suggestions to be pursued, but I do think the paper would be substantially improved if this main point can be substantially addressed.*

We agree that a clear distinction from previous work (especially previous life cycle work) is very important. We think that the changes in the linear growth phase observed in the experiments of Wittman et al. and Smy and Scott can be explained by differences in tropopause-level PV gradients and tropospheric winds between their experiments with and without polar vortex, as was already noted in Section 3.2. A direct change in tropopause-level PV gradients (imposed by the initial conditions) leads to modifications of growth phase of the life cycles (e.g., a change in conversion from APE to EKE via the eddy heat flux). In our experiments we tried to specifically minimise this direct stratospheric impact, which can be difficult to distinguish from sensitivities in a purely tropospheric setup. To emphasise this and other major differences in results extend existing notes and added additional notes about the different basic states used in the respective studies at various places in the manuscript (lines 76-78, 95-100, 201-206, 387-394) to further emphasise that our setup has almost no change in tropopause-level PV gradient in contrast to the setups used in other studies and we correspondingly do not observe any changes to the linear growth phase.

### **Major comments/suggestions**

#### **PV structure**

*The PV structure of the initial conditions is emphasized in Fig. 1, but it is not really made reference to later in the interpretation. For instance, how do the perturbations considered in Section 4 modify the PV gradients? One clear difference between T and TS is the presence of narrow region of positive pv gradient near 20 km at 60 degrees in the presence of a stratospheric jet; is it really the winds in the lower stratosphere that are key, or is it the PV gradient structure? Moreover, how is the zonal PV structure changed over the evolution of the lifecycles?*

The main reason for displaying the PV gradients in Fig. 1 was to draw attention to the regions of negative gradient that commonly appear in the initial conditions of typical life cycle experiments, but in our experience they are often not reported (although we concluded that those regions have no major effect on the evolution of the life cycle).

The PV distribution (and in particular the PV gradient) are certainly very important in determining the dynamics of the system. However, especially the zonal structure of the response is strongly dictated by the (non-linear) evolution during the decay phase of the life cycle (since the initial state has no zonal structure) and it is not trivial to make statements about the precise mechanisms or dynamical features of the initial condition setup and processes that lead to the observed NAM signal.

Throughout the manuscript we describe most changes in initial state or life cycle evolution via the wind field (since the system is defined via the wind field), but tried to make it clear that the (structure of the) wind of course couples to various other characteristics of the system, like the temperature and PV fields. Part of the evolution (in particular the zonal structure) is described in Fig. 2 in terms of PV.

During our analysis of the system we did not find that using the PV field to be more useful than the wind field to describe the final state of the life cycle (and the corresponding dependency on stratospheric dynamics) and thus we do not think that adding PV diagnostics to the analysis will improve the manuscript. Further, finding crucial differences in PV between the experiments can be difficult since the PV field tends to become rather noisy during the non-linear phase (compared to the wind field), which is even more the case for the PV gradient (as it essentially scales with the curvature of the wind). However, to emphasise the coupling between PV and the wind field and due to general importance of PV (gradient) for the system we added corresponding note (line 219).

## **Energetics**

*It seems that one way of interpreting these results, consistent with the discussion of Barnes and Young (1992) in Section 1.2, is that the presence of the stratospheric jet allows for a greater conversion of APE to EKE and ultimately MKE; and that this is further enhanced by including surface friction. If this is a fair characterization of the present results, how does the available potential energy differ in the various cases? Could this result be simply explained by the presence of more APE in the cases with a stratospheric vortex?*

We find that initial state APE is indeed increased in our experiments with stratospheric jet and ultimately the APE to MKE conversion is enhanced in life cycles with stratospheric jet (consistent with the observed jet shift). However, since this increase in APE corresponds to a change of the temperature structure in the stratosphere it is not obvious that and how the additional stratospheric APE will be converted into tropospheric MKE during the life cycle. Further, we do not find a change in life cycle during the linear growth phase (where APE to EKE conversion is dominant), suggesting the life cycle to be unaffected by the direct increase of initial APE. To address these issues extended our discussion of the global energetics and included a note regarding changes in initial state APE and corresponding APE to MKE conversion (lines 411-419).

*A related comment: in the presence of surface friction there isn't really final steady state as MKE will continue to dissipate. This is very briefly discussed right at the end of the discussion (I430), but until the reader gets there it's not really clear how the authors defined the final state. Would it be possible to put the figures in the supplementary information on more equal footing by computing the energy lost to surface friction in the runs in Fig. S2, showing instead the net EKE and MKE generated by conservative processes?*

We added a note just after the introduction of the first experiments including surface friction (Section 3.3, lines 286-289) explaining that we define a 'final state' analogous to the case without friction for easy comparison of the two cases. We further (as also mentioned above) extended our discussion of the global energetics in the discussion section and the supplementary material in order to address the issue of comparability of experiments with and without surface friction, including added discussions (and Figs. S2 and S3 in the supplement) regarding barotropic energy conversion (lines 260-262, 411-419, 502-504 and supplement).

### **Mass fluxes**

*There is some discussion of the 1000 hPa 'Geopotential height' anomalies as a measure of the surface NAM or AO signature of the baroclinic adjustment. One substantial point is that equation (2) on line 270 assumes gradient wind balance holds near surface. This is ok for case with no friction, but not for the case with surface friction, which will modify dominant the force balance. This could be corrected by adding in surface friction on right hand side. But I wonder if it might be more directly useful to consider the evolution of the surface pressure, and the nature of the mass fluxes in these experiments; in particular I wonder if these are substantially modified by the presence of an Ekman layer in the cases with surface friction.*

As the model equations are formulated in absolute pressure surface pressure is not a model variable and corresponding diagnostics are not available. The lack of surface pressure data further does not allow for a direct computation of the geopotential via vertical integration of temperature, which is why we decided to use a definition of geopotential based on the wind field. The geopotential field (on 1000hPa) seems to be a suitable diagnostic to analyse in regards of AO or NAM like phenomena, as is it typically used to define the corresponding indices. We agree that Equation 2 is missing a friction term. However, for a friction time scale of 1 day the missing term ( $-a \cdot \bar{v} / \tau$ ) is typically small compared to the Coriolis term ( $-a \cdot f \cdot \bar{u}$ ). We found a difference in magnitude of the terms of about a factor 100 for the final state of our experiments. We added a note next to Equation 2 to clarify that we neglect the friction term (lines 310-311).

### **Specific/Minor Comments:**

*I76: perturbation to what?*

We added a 'stratospheric PV' before perturbation (lines 87-88).

*I188: could this alternatively be explained by greater meridional shear in TS simulation?*

The shear would probably affect wave behaviour via changes in PV gradient, as mentioned. We added a note to link the PV gradient to the wind structure again as it is probably important to emphasise this connection throughout the paper (line 219).

*I213: Are EKE and MKE integrated over depth of troposphere or the depth of atmosphere? I would think that the MKE of polar jet is small but not negligible.*

Energies are integrated over the entire atmosphere (NH only). The difference in MKE between initial states T and TS is only a small fraction of the MKE difference in their final states. Further, the difference in final state MKE mostly persists when integrating only up to

100hPa, i.e., when we exclude the stratospheric jet from the energy computation. In order to subtract the minor contribution of the stratospheric jet to the MKE, we show the difference to the initial state rather than full MKE. We modified the corresponding part of the manuscript to clarify how the energy densities are calculated (line 242).

*l310-317: I found the grouping of these experiments to be initially quite confusing, thinking that the natural ordering would be to group  $TS > 10$  with  $TS < 10$ , and  $TS > 25$  with  $TS < 25$ . I did ultimately figure out that the logic is intended to emphasize the wind anomalies in the lower stratosphere (to be fair, this is stated in the text, but preconceived notions can be hard to shake sometimes).*

*This is ultimately just an issue of presentation. But I did wonder if it would be clearer to readers to start with (what seems to me) the more natural a priori sorting instead of anticipating the results? If not, perhaps naming the sets something more descriptive, such as 'Weak winds/Strong winds' instead of 'Set 1/Set 2'?*

We changed the names of the groups accordingly for more intuitive understanding.

*l325 discussed*

Corrected.

*Fig. 7: How does  $u$  at 15 km vary over this set of integrations? This might be more illuminating than the mid-stratospheric winds.*

Fig 7 shows final state zonal wind at 10km, which is where we find the strongest response. Winds at 15km change qualitatively the same throughout the simulation as also suggested by the rather barotropic final state winds shown in Fig. 5. Initial state winds at any height change linearly with stratospheric jet magnitude  $u_{Smax}$  due to the stratospheric jet being superimposed onto the tropospheric jet. No corresponding changes to the manuscript were made.

*l413-414: what is this statement based on? explicit calculations or assumption of linearity?*

As mentioned, we performed a set of sensitivity experiments with increased tropospheric jet magnitude (and no stratospheric jet) what showed a similar NAM response as the experiments with stratospheric jet, but the required difference in tropospheric jet magnitude was much larger than the increase induced by inclusion of a stratospheric jet in order to obtain a NAM response of similar magnitude. This suggests that the increase in wind speed at tropopause level alone is not sufficient to explain the observed NAM response in our experiments. We changed the corresponding part to '...it was necessary to increase the jet magnitude by order of 10 m/s in these sensitivity experiments...' in order to clarify that the statement is based on explicit simulations (lines 481-482).

*l423: The additivity of surface winds also is not perfect.*

We added a corresponding note to the paragraph in order to slightly weaken the statement made (line 492).

*l450: This final paragraph also contributed to my feeling unsatisfied by the discussion. What is the explanation for the downward influence suggested by the present results? How can we use these experiments to quantify the eddy feedback? These aren't unreasonable claims, but the authors should explain them more explicitly.*

We agree that a concise, yet complete and clear, description of our main findings and arguments in Section 6 is important and therefore extended the respective last paragraph, as suggested by the referee (lines 533-541).

## Response to Referee 2:

*In this manuscript, the authors use an idealized model to investigate the role of the stratospheric polar vortex on influencing the tropospheric jet. Specifically, they test the sensitivity of the NAM-like equatorward jet shift associated with a weak or reversed stratospheric polar vortex to the height of the vertical winds in the polar jet, and to the magnitude of surface friction. They find that the tropospheric response is sensitive to changes in winds in the lower stratospheric polar vortex, but not to winds in the upper stratosphere. Additionally, they find that surface friction enhances the tropospheric response to stratospheric vortex changes, and that friction acts to bring nearly barotropic anomalies all the way down to the surface.*

*Although the findings in general agree with previous studies, the consolidation of the impacts of surface friction and lower-stratospheric anomalies using a coherent model framework produces a compelling standalone study with implications for our dynamical understanding of sudden stratospheric warmings. The manuscript is well-written, well-organized, and the scientific questions well-constructed, and I only have a couple minor comments and questions. Most of my corrections and minor questions can be found in the attached manuscript.*

*The main additional question I have is if the authors think this study can provide insight into the observed differences in timing between "Displacement" type sudden stratospheric warming events, and "Split" type sudden stratospheric warming events. Specifically, Splits are observed to show an almost-instantaneous NAM-like response to SSWs, whereas for Displacements, the response tends to occur with a significant lag on the order of weeks. Could this be related to a difference in the vertical structure of the polar vortex in Displacement versus Split events? For example, is it possible that Displacement events show anomalies that start higher up in the stratosphere and work their way downwards, while Split events produce a strong signal in the lower stratosphere almost immediately? I am not sure why or how surface friction could play a role, since I can't imagine any way in which surface friction could depend on the type of vortex breakdown.*

The idea of obtaining insights into differences between displacement and split events is surely interesting. Smy and Scott (2009) perform life cycle experiments with initial conditions modelling a polar vortex in a split or displacement state based on an imposed PV field and a corresponding PV inversion.

Directly inferring results from the present experiments regarding different SSW states might be a bit speculative and go beyond the scope of this study. However, we think the described set-up and methodology is in general suitable to study such problems and we intend to perform life cycle experiments with more realistic initial conditions (including split and displacement states) at some point. We added a paragraph in the discussion about potential extensions of this study (lines 505-515).

*Line 155: Should we expect the influence of the stratospheric jet on the tropospheric circulation to peak for tropospheric zonal wave numbers of 6 and 7? Does it have something to do with the relative baroclinic instability for each zonal wave number?*

The strength of the baroclinic life cycle (in our setup) is strongest for wave numbers 6 and 7. When using different wave number perturbations (e.g. 5) the life cycle is in general weaker and therefore the absolute anomaly induced by the inclusion of stratospheric winds is expected to be weaker, too. We extended the note regarding this wave number sensitivity in our manuscript (lines 178-181).

Various typos have been corrected based on the notes within the supplementary material provided by the referee.



## Response to Referee 3:

*This paper posed an interesting scientific question regarding stratosphere-troposphere coupling, and addressed it with well-designed experiments. I think the authors' use of a more simplified atmospheric model was a good choice here, as these simplified simulations are able to capture most of the atmospheric dynamics relevant to their scientific interests. The experiments described in this paper explore the tropospheric response to the presence or absence of a stratospheric jet, in an idealized attempt to compare conditions that typically occur in wintertime (a strong stratospheric jet), and conditions that are common after a sudden stratospheric warming event (a much weaker or absent stratospheric jet). The authors explore a robust parameter space in their experiments, and they analyze a variety of stratospheric jet strengths, several different vertical structures of stratospheric winds, and the effects of surface friction. The authors find that when a stratospheric jet is absent, the tropospheric jet is located equatorward of where it is located in the presence of a stratospheric jet. The presence of surface friction enhances this response, with an even stronger equatorward shift of the tropospheric jet when surface friction is present and a stratospheric jet is absent. Finally, the authors note that the winds in the lower stratosphere are key for influencing the tropospheric response; when the wind anomalies are present only in the middle or upper stratosphere, the tropospheric response is weak.*

*Overall, I found the scientific question in this paper to be well-posed, the experiments to be well-designed, the authors' interpretation of results to be logical and interesting, and the paper to be well-organized and well-written. This manuscript is appropriate for publication in Weather and Climate Dynamics, and would make a good contribution to the scientific community. I recommend that this paper be Accepted with Minor Revisions. I have a few minor comments for the authors, and a few small suggestions that I think could strengthen this manuscript further, but I do not need to see this manuscript again. My comments and suggestions are presented in line-by-line format below.*

### **Minor Comments**

1. L17: "The troposphere and the stratosphere form a dynamically coupled system."

[The text was corrected accordingly.](#)

2. L20: Remove the word "maybe"; you could replace it with "Some of the most prominent stratospheric phenomena..."

[The text was corrected accordingly.](#)

3. L31-32: Replace "it can lead to periods with weak and equatorward shifted tropospheric jet stream" with "a polar vortex break-down can lead to periods with a weak and equatorward shifted tropospheric jet stream".

[The text was corrected accordingly.](#)

4. L66-68:

5. L77: Your left quotation mark around "sub-vortex region" is backwards; if you're using LaTeX for your manuscript, you probably need to use the ``` key instead of the `'` key.

[The text was corrected accordingly.](#)

6. L89: Left quotation mark around "secondary cycles" is backwards; see above.

[The text was corrected accordingly.](#)

7. L145: Change "configuration" to "configurations".

The text was corrected accordingly.

8. L146: Change "(slightly)" to "( a slightly)".

The text was corrected accordingly.

9. Lines 154-157: *When I think about stratosphere-troposphere interactions in the context of SSWs, I often think about the role of planetary-scale waves (e.g., zonal wavenumbers 1-3) rather than the smaller-scale waves in your experiments. Since your results do seem to show some sensitivity to wavenumber, do you think that larger waves would respond to your stratospheric perturbations in a similar way? Or do you think the behavior would be completely different?*

We agree that planetary waves can play an important role in downward coupling following SSWs. However, our experiments are specifically set up to isolate and study the role of synoptic waves related to baroclinic instability in the troposphere (related to tropospheric eddy feedback). In that sense we are not in a position to say much about the role of planetary waves based on our experiments, which are not primarily controlled by baroclinic instability but rather by forcing due to flow over topography etc. We included a short remark clarifying this distinction (lines 178-181).

10. Lines 177-178: *I found the sentence starting with "However, especially the non-linear decay phase..." a bit confusing. Is this sentence trying to state that wave growth (first few days) is not substantially influenced by the presence of a stratospheric jet? If so, it might help to say that more explicitly. Also, is this result in opposition to some of the results you discuss in your introduction? My understanding is that Wittman et al. (2007) and perhaps Smy and Scott (2009) both saw changes in the growth rates of the baroclinic waves in their experiments, which seems to me to be different to what you saw. This could be a good place to refer back to these earlier works, and perhaps try to explain the discrepancies a bit. I realize you make this comparison a bit later in the paper (around lines 215-220), but I would consider moving this comparison a bit earlier.*

We agree that this would be a useful place to first mention the vanishing influence of stratospheric winds on the linear phase, a corresponding note and reference were added to the paragraph (lines 201-206), but we left the main discussion of this issue where it is (within the paragraphs describing Fig. 3). We further added notes throughout the manuscript emphasising the distinction between our findings regarding the linear growth phase and the findings of other authors (95-100, 201-206, 387-394).

11. Line 196, 198: Again, backwards quotation mark around "secondary cycles".

The text was corrected accordingly.

12. Figure 4: *Overall, your figures were quite good; however, this figure was a bit confusing for me. We are comparing panels b-d directly to panel a, correct? That is, the shading indicates the difference between the black lines in panels b-d and the black lines in panel a? It might help to say that very explicitly in the caption. I don't think you need to change your figures necessarily (though adding a vertical line at day 6 could be nice but is not necessary), but I would suggest being very direct about what we're looking at, because it took me a couple of minutes to properly orient myself (and you have several other figures that follow this convention, so being painfully obvious about it the first time might be helpful). I also would suggest you might want to flip how you describe Figure 4--when I look at panels b-d in Figure 4, the first thing my brain thinks is "equatorward shift". But in the discussion beginning in Line 240, you talk about a poleward shift first, which took me a minute to comprehend. So maybe it would help to mention the equatorward shift without the stratospheric jet BEFORE you mention the poleward shift with the stratospheric jet. Again I don't think you need to change anything about the figure, just the order in which you discuss things.*

As suggested by the referee we extended the caption and the describing paragraphs of Fig. 4 to simplify the interpretation of our results (lines 270-273). We further added a visual aid to the figure to help identify day 6.

13. Discussion in Section 4 : *When I was reading this section, I thought back to some of the experiments of Butler et al. (2010). The authors imposed a polar stratospheric cooling anomaly in a simplified model (not winds directly, as you do, but a stronger stratospheric jet was produced in response to the cooling). They tested several different heights of their temperature (and thus, wind) anomalies, and their results seem very complementary to yours--when the temperature anomaly was limited to the middle or upper stratosphere only, the tropospheric response was weaker or non-existent. I'd encourage you to check out their results (centered around their Figure 5), as I think this strengthens some of the points you're making in this section. To that end, zonal mean temperature anomalies for your experiments could be interesting, if you have them. My thinking is that it could be nice to more seamlessly link your work to some of the other idealized modeling work that simulates polar stratospheric variability with a temperature anomaly, instead of a wind anomaly (think Figure 4 but for temperature). This is just a thought though--I don't think it would really fit in well in your main manuscript but they could be a good supplemental figure, or even just for your own knowledge.*

The connection to Butler et al. (2010) is indeed interesting and insightful, further emphasising the range of studies and experimental setups that show a tropospheric jet-shift response to (direct or indirect) wind perturbations in the lower stratosphere. We added references to Butler et al. to the introduction and an additional note to the discussion (lines 47-51, 429-431, 468-469).

14. Line 314-15: Backwards quotation marks around "Set 1" and "Set 2".

The text was corrected accordingly.

15. Line 333: Remove " does " and change "increase" to "increases ".

The text was corrected accordingly.

16. Line 364: Change "worth to point out" to " worth pointing out ".

The text will be corrected accordingly.

17. Lines 365-367: Butler et al. (2010) (and maybe Polvani and Kushner (2002)) also show increases in eddy momentum fluxes near the tropopause in response to polar stratospheric cooling (and an increase of the winds, similar to your TS), so you could cite that here as well if you'd like.

We agree that it would be beneficial for the reader to be reminded of the mentioned studies; references were added to this paragraph as suggested (lines 429-431).

18. L396: Change "weather" to " whether ".

The text was corrected accordingly.

19. Lines 409-418: Another thing that occurred to me during your discussion of troposphere-only jet variability is that the jet response in many simplified and comprehensive models is dependent on its initial state--that is, jets that start farther equatorward tend to shift more in response to the same perturbation as a jet that starts closer to the pole (e.g., Barnes et al. (2010), Kidston and Gerber (2010)). Furthermore, the presence or absence of the stratosphere itself can have a large impact on the initial state of the jet (e.g., Wang et al. (2012)). Since the perturbations that lead to SSW events often come the troposphere, I suppose the point I am trying to make is that the background state can substantially modify the tropospheric jet response to atmospheric perturbations, and the background state of the troposphere can perturb the vortex in the first place. So at some point when thinking about SSWs it's necessary



*to zoom out and think about that troposphere-to-stratosphere pathway (not saying you need to change your analysis at all, but I think it is worth mentioning in the discussion at some point).*

We agree that it is important to keep in mind that troposphere and stratosphere form a system with coupling in both directions and therefore any climate forcing modifying the tropospheric jet will also affect the stratospheric jet, which can then feed-back again onto the tropospheric jet. We added a note to the discussion section to remind the reader about this fact (lines 424-425).

20. Line 445: Change "fiction" to " friction ".

The text will be corrected accordingly.

21. Lines 450-455: *Is your explanation for the downward control essentially that changes in lower stratospheric winds (specifically, a weakening of the lower stratospheric winds on the flank of the polar vortex as generally occurs in response to a strong SSW event) drives a negative annular mode-like response in the troposphere? I'd recommend saying that very clearly and explicitly, since this is your last opportunity to concisely summarize your argument.*

We agree that a concise, yet complete and clear, description of our main findings and arguments in Section 6 is important and therefore extended the respective last paragraph, as suggested by the referee (lines 533-541).

# Tropospheric eddy feedback to different stratospheric conditions in idealised baroclinic life cycles

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## Abstract.

A pronounced signature of stratosphere-troposphere coupling is a robust negative anomaly in the surface northern annular mode (NAM) following ~~major~~ sudden stratospheric warming (SSW) events, consistent with an equatorward shift of the tropospheric jet. It has previously been pointed out that tropospheric synoptic-scale eddy feedbacks, mainly induced by anomalies in the lowermost extratropical stratosphere, play an important role in creating this surface NAM-signal. ~~We~~ Here, we use the basic setup of idealised baroclinic life cycles to investigate the influence of stratospheric conditions on the behaviour of tropospheric synoptic-scale eddies. Particular focus is ~~hereby~~ given on the enhancement of the tropospheric eddy response by surface friction, as well as the sensitivity to wind anomalies in the lower stratosphere. We find systems that include a tropospheric jet only (modelling post-SSW conditions) to be characterised by an equatorward shift of the tropospheric jet in the final state of the life cycle, relative to systems that include a representation of the polar vortex (mimicking more undisturbed stratospheric winter-time conditions), consistent with the observed NAM-response after SSWs. The corresponding relative surface NAM-signal is increased if the system includes surface friction, presumably ~~associated with~~ due to a direct coupling of the eddy field at tropopause level to the surface winds. We further show that the jet shift signal observed in our experiments is mainly caused by changes in the zonal wind structure of the lowermost stratosphere, while changes in the wind structure of the middle and upper stratosphere have almost no influence.

## 1 Introduction

### 1.1 General background

~~Troposphere~~ The troposphere and stratosphere form a dynamically coupled system. In order to better understand tropospheric weather and climate behaviour it is essential to understand how stratospheric conditions and processes can have a downward influence and modify the tropospheric circulation or produce surface signals.

~~The maybe~~ Some of the most prominent stratospheric phenomena in the northern hemisphere are (major) sudden stratospheric warming (SSW) events. During these sudden warmings the westerly winds of the stratospheric polar jet (also polar vortex) break down or even reverse. Thompson and Wallace (1998) showed that the winter time variability of the stratospheric polar vortex and the tropospheric mid-latitude jet are strongly correlated.

25 Baldwin and Dunkerton (2001) used a composite study of weak vortex events (of which SSWs would form the extreme subset) to investigate the time dependence of this coupling in more detail. They showed how the stratospheric zonal wind anomalies propagate downwards into the troposphere and demonstrated that this downward influence appears to have two components: at first ~~the signal modifies the wind structure above and at tropopause level on sub-seasonal to seasonal time scales, creating a long-lasting zonal wind anomaly in zonal wind anomalies extend downward through~~ the lower stratosphere  
30 ~~. From there the signal where they can last for several weeks. Once the signal reaches the tropopause level it~~ can penetrate quasi-instantaneously into the troposphere and create ~~surfaces~~ surface anomalies that persist on weekly time scales.

Since then various studies have supported the idea that the break-down of the polar vortex can have a downward impact and induce zonal wind anomalies in the troposphere. In particular, ~~it a~~ a polar vortex break-down can lead to periods with weak and equatorward shifted tropospheric jet stream. This equatorward shift of the jet typically manifests as negative anomaly of the  
35 northern annular mode (NAM) index or similar indices (~~e.g., Karpechko et al., 2017; Charlton-Perez et al., 2018~~) (e.g., Thompson and Wallace, 1998; Karpechko et al., 2017; Charlton-Perez et al., 2018). Changes in the large scale tropospheric circulation can then affect local surface weather and, with it, change the likelihood of extreme events like cold spells (Thompson and Wallace, 2001; Kolstad et al., 2010; Kautz et al., 2020).

Different mechanisms have been proposed to explain the downward propagation of stratospheric wind anomalies and their  
40 influence on the tropospheric circulation. However, no single fully conclusive mechanism has been ~~found~~ identified yet. Note that, in addition, the tropospheric response to SSWs might also be caused by a combination of different coupling processes. One of these potential coupling processes is given by tropospheric eddy feedback as a response to the induced stratospheric anomalies. Domeisen et al. (2013) have shown in idealised model runs that tropospheric ~~eddies are~~ eddy feedback is essential to obtain a robust negative NAM signal following a SSW. Hitchcock and Simpson (2014) also found tropospheric synoptic-scale  
45 eddy feedback to play a significant role in creating a NAM-like surface response. They further concluded that the most relevant aspect of the stratospheric variability does not seem to be the wind reversal in the mid-stratosphere, but the persistent wind anomalies in the lowermost stratosphere. Butler et al. (2010) performed a series of steadily forced idealised model experiments with imposed cooling either in the entire polar stratosphere or confined to the middle and upper polar stratosphere (mimicking, e.g., ozone-hole induced climate change), directly causing a consistent wind anomaly in the respective region. They found the  
50 troposphere to respond to the imposed stratospheric anomalies in a NAM-like fashion if the imposed anomalies reach into the lower stratosphere, but the tropospheric response to be weak if the anomalies are confined to the upper stratosphere. Karpechko et al. (2017) showed that in both, model runs and reanalysis data, SSWs which produce strong and long-lasting anomalies in the lowermost stratosphere have an increased likelihood for a tropospheric impact compared to SSWs with weak anomalies in the lowermost stratosphere.

## 55 1.2 Previous baroclinic life cycle work relevant for this study

A simple, yet fundamental, way to investigate the role of synoptic scale eddies in the dynamical coupling between stratosphere and troposphere is through (idealised) baroclinic life cycle experiments, an initial value problem starting from an imposed baroclinically unstable tropospheric jet. During the subsequent break-down of the imposed jet a baroclinic wave can be observed to

develop, grow and eventually decay, leaving the system in a state with a more barotropic, strengthened and poleward shifted jet compared to the initial conditions (see, e.g., Simmons and Hoskins, 1978; Thorncroft et al., 1993). Such life cycle experiments have previously been used to study the influence of stratospheric winds onto the evolution of tropospheric baroclinic eddies.

Wittman et al. (2004) performed idealised life cycle experiments using initial conditions that either do or do not include winds in the stratosphere, representing situations with an intact or a broken-down polar vortex. They found that if the system includes a polar vortex the evolution of the life cycle is strongly modified and when the polar vortex is removed the system exhibits a (weak) dipole structure in the surface geopotential height field, similar to the surface NAM response observed after SSWs, which corresponds to an equatorward shift of the tropospheric jet. They further note that this surface signal is weak if the polar vortex is rather confined to the stratosphere, but gets strongly enhanced if the polar vortex reaches deep into the troposphere.

In a following study Wittman et al. (2007) investigated the role of stratospheric vertical shear onto the evolution of baroclinic life cycles. They used three different setups in which the winds of the tropospheric jet either decreased, stayed constant or (further) increased above the jet core. For the three situations they found pronounced differences in the evolution of the life cycle, including substantial changes in the growth rate of the baroclinic waves and the qualitative characteristics of the wave growth and decay phases. It should be noted, that the initial conditions used by Wittman et al. (2007) were mostly motivated to resemble a setup of the Eady model for baroclinic instability, rather than realistic atmospheric conditions. The corresponding change of stratospheric shear induces strong changes in the vertical curvature of zonal wind at tropopause level, and thus strong changes in the meridional gradient of potential vorticity (PV) in that region, which are known to have a strong impact on the evolution of baroclinic waves in the troposphere. In the present study we specifically design initial conditions that do not substantially modify tropopause-level PV gradients to minimise their direct impact on the development of baroclinic instability.

Kunz et al. (2009) used a similar setup as Wittman et al. (2004) and also found that the presence of a stratospheric jet can qualitatively alter the evolution of the baroclinic life cycle. Further, they could not explain the modification of the life cycle with simple refractive index linear theory and therefore concluded that the non-linear part of the wave evolution plays an important role in the coupling.

Smy and Scott (2009) investigated the influence of stratospheric PV anomalies on the evolution of idealised baroclinic life cycles to obtain insights into the dynamical coupling of stratosphere and troposphere during and after SSWs (including a distinction between split and displacement events). They reported a decrease in growth rates and general wave activity (and a corresponding reduction in magnitude of the surface geopotential anomaly of the final state) with increasing stratospheric PV perturbation amplitude. ~~Smy and Scott (2009) further comment on the influence of the 'sub-vortex region', given by the lowermost extratropical stratosphere. A modification of the wind structure (or equivalently the PV field) in this region can represent the direct effect of stratospheric anomalies on the tropospheric winds. They found that the influence of the polar vortex on the life cycle evolution decreases as the stratospheric jet reaches deeper into the lowermost stratosphere.~~ These results partly seem to be in disagreement with the results of Wittman et al. (2007) or Kunz et al. (2009). However, Smy and Scott (2009) also note that their results might be explained by a change in tropospheric horizontal shear due to the non-local effects

of the stratospheric PV anomaly and a corresponding fundamental change in the nature of the life cycle (see also Thorncroft et al. (1993) for details on how horizontal shear can affect the evolution of baroclinic waves). [Smy and Scott \(2009\)](#) further comment on the influence of the ‘sub-vortex region’, defined as a region in the lowermost extratropical stratosphere without imposed stratospheric PV anomaly in their experiments and therefore reduced strength of the lower stratospheric winds. A modification of the wind structure (or equivalently the PV field) in this region can represent the direct effect of stratospheric anomalies on the tropospheric winds. They found that the influence of the polar vortex on the life cycle evolution increases as the stratospheric jet reaches deeper into the lowermost stratosphere.

While much focus was given on sensitivities of the linear growth phase of baroclinic life cycles to various changes of the system Barnes and Young (1992) also investigated the evolution during the non-linear decay phase to a range of flow-dependent forcing processes, including surface friction. They found the system to undergo a series of growth and decay phases in cases with sufficiently weak diffusion, in contrast to the single growth phase with subsequent decay of eddy energy in cases with strong diffusion. They further showed that simulations with surface friction can produce more pronounced such ‘secondary cycles’, i.e., growth and decay phases following the initial life cycle, as the surface drag tends to work against the barotropisation of the non-linear phase and thus act as source of baroclinicity.

### 1.3 Potential influence of surface friction

The influence of surface friction onto the evolution of baroclinic eddies is potentially crucial to understand the surface signal observed after SSWs, as it can be argued that the inclusion of surface friction increases the potential for the mid- and upper-tropospheric eddy field to couple to the surface winds. This can be illustrated using [an approximated version of](#) the evolution equation of the vertically averaged zonal mean zonal wind, given in Equation 1 (see, e.g., chapter 10 of Vallis (2017)).

$$\partial_t[\bar{u}] = -\partial_y [\overline{u'v'}] - \bar{u}_{sf}/\tau, \quad (1)$$

where  $u$  and  $v$  are zonal and meridional wind,  $\bar{u}_{sf}$  the zonal mean zonal surface wind,  $\tau$  the surface friction time scale, square brackets and overbars denote vertical and zonal averages, respectively, and primed quantities describe deviations from the zonal mean (note that we neglected the mean flux term as it tends to be small in our system, consistent with quasi-geostrophic scaling). Here we used a linear damping of surface winds as simple parametrisation of surface friction. In the case with vanishing friction ( $\tau \rightarrow \infty$ ), only the meridional momentum fluxes can act as source for (vertically averaged) zonal momentum and changes in  $\bar{u}$  tend to occur in regions of non-zero momentum flux, i.e., around tropopause level for baroclinic life cycle experiments. For finite values of  $\tau$ , on the other hand, the atmosphere can ‘exchange’ momentum with the surface, allowing for a non-local coupling between surface winds and the eddy field. This additional coupling mechanism suggests that a dynamic modification of the eddy field (due to the presence of a stratospheric jet) can lead to an enhanced change of the corresponding surface winds (in terms of the difference between final and initial state) in cases where surface friction is active in the system.



In the present paper we further investigate what impact the presence of a stratospheric polar vortex has on the idealised tropospheric baroclinic life cycle. In particular we are interested in the sensitivity of the life cycle evolution to changes in wind structure in the lower stratosphere, compared to changes in the middle and upper stratosphere, and the influence of surface friction onto the surface signal of the life cycle induced by the presence of a stratospheric jet. We hereby mostly focus on the modification of the equilibrated ~~the~~ ‘final’ state of the system, as opposed to the details of the (linear) growth stage or the (non-linear) decay stage of the baroclinic wave.

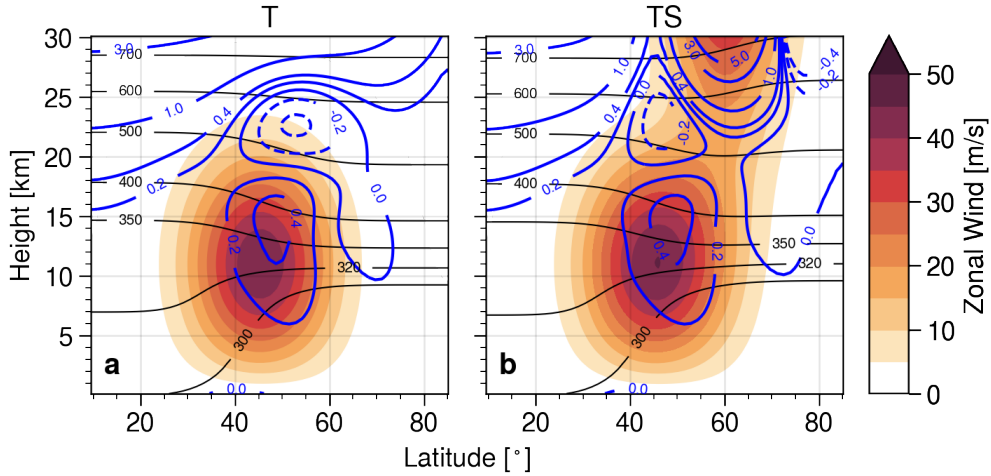
Section 2 introduces the model setup used in this study and lays out the specifics of the different sets of initial conditions. In Section 3 we discuss in detail various changes of the evolution of the baroclinic life cycle due to the presence of stratospheric jet, with particular focus on the NAM-like response of the troposphere in the final state of a life cycle when there is no stratospheric jet present, compared to when there is. Additionally we show that we only find a strong signature in the corresponding surface signal when the system is subject to surface friction. We then provide evidence, in Section 4, to show that this NAM-like signal is mainly caused by the modification of winds in the (extra-tropical) lower stratosphere and the inclusion of winds in the middle and upper stratosphere have almost no influence on the final state of the life cycle. In Section 5 we further discuss and interpret some of our findings before, in Section 6, we summarise the main conclusions of this paper.

## 140 **2 Model and Basic states**

All simulations are run with the simple dry dynamical core model BOB (Built on Beowolf, see Rivier et al. (2002) for details). The model solves a spectral representation of the primitive equations in pressure coordinates with truncation at horizontal wave number 85. The discrete vertical levels are distributed with constant spacing  $\Delta z = 250$  m up to a height of  $z = 60$  km, where  $z = -H \ln(p/p_0)$  is a log-pressure coordinate with scale height  $H = 7.5$  km and reference pressure  $p_0 = 1000$  hPa. To minimise upper boundary effects we add 10 additional model levels between  $z = 60$  km and  $z = 82$  km, equally spaced in pressure. Note that we are using a substantially higher vertical resolution than has typically been used in similar studies, since we found in particular the details of the non-linear decay phase of the baroclinic life cycles to be sensitive to changes in  $\Delta z$  for values larger than about  $\Delta z = 250$  m, as also further explained in Section 3.

The model is initialised with a prescribed state and integrated forward in time with a step length of 5 minutes over a period of 30 days, giving daily output of instantaneous fields (results are qualitatively unchanged for hourly output). ~~We initialise all experiments~~ To ensure numerical stability and model energy dissipation via sub-grid-scale processes the model includes an 8th order hyper-diffusion, damping the smallest resolved wave numbers on a time scale of 2.4 hours (for comparison Wittman et al. (2007) use a 6th order diffusion with 1 hour time scale; Kunz et al. (2009) use an 8th order diffusion with 6 hours time scale at T42 resolution).

All experiments are initialised with an idealised and zonally symmetric basic state, loosely based on the initial state used by Kunz et al. (2009). The basic state is analytically defined via a given zonal wind field and is chosen to represent two general situations, depending on the choice of parameters: either a system with a tropospheric jet only (modelling post-SSW condi-



**Figure 1.** Examples of the basic state used in this study with different choice for parameter values to include either a) a tropospheric jet only (experiment T) or b) a tropospheric and a stratospheric jet (experiment TS). The shading shows the zonal wind, thin black contours show potential temperature [m/s/K] and thick blue contours show the meridional PV gradient [PVU/deg], with dashed contours corresponding to negative values.

tions), or a system that contains a tropospheric and a stratospheric jet (mimicking more undisturbed winter-time conditions). In order to also study the sensitivity of changes in the wind structure of different regions in atmosphere we further use a set of basic states which include the tropospheric jet and only the upper or lower part of the stratospheric jet, respectively. Table 1 summarises the different types of basic state configurations used in the present study. The two main basic state configurations (T and TS) are visualised in Figure 1 (note that only a part of the domain is shown).

**Table 1.** Different basic state configurations used.

Experiment	Description
T	Tropospheric jet only
TS	Tropospheric and stratospheric jet (with magnitude $u_{Smax} = 75\text{m/s}$ )
$TS_{<z_\eta}$	Tropospheric and lower part of the stratospheric jet (below height $z_\eta$ )
$TS_{>z_\eta}$	Tropospheric and upper part of the stratospheric jet (above height $z_\eta$ )

The temperature distribution of the respective initial state is calculated to be in thermal-wind balance with the prescribed wind field. Note that the resulting meridional PV gradient (thick blue contours in Figure 1) strongly depends on the vertical curvature of the underlying wind field and therefore produces a pronounced local maximum near the tropospheric jet core.

Further note that both configurations displayed in Figure 1, due to the strong dependency on the wind field structure, include regions with (a slightly) negative PV gradient, which could potentially influence the evolution of the life

cycle. However, the corresponding initial states follow the typical setup used in this type of idealised life cycle experiment. We further performed a series of sensitivity experiments and concluded the regions of negative PV gradient to have no significant influence on the qualitative results presented in this paper. Magnusdottir and Haynes (1996) also raised the question of the effect of negative PV gradients in typical life cycle setups on the evolution of the baroclinic wave and concluded that these regions can have an effect on certain details of the non-linear phase (e.g., details of the energetics), but seem to have no impact on most aspects of the qualitative behaviour.

To trigger the growth of a baroclinic wave the initial state is perturbed by super-imposing a zonally periodic near-surface temperature perturbation of fixed zonal wave number 6, centred around  $45^\circ$  latitude. We found our results to be qualitatively similar for perturbations with wave number 7, but the stratospheric jet to have almost no influence on the life cycle for wave numbers 5 and 8 (in these cases the purely tropospheric life cycle is generally weaker than for perturbations with wave numbers 6 and 7). Our main focus is to study the behaviour of tropospheric waves at synoptic scales, which are primarily forced as a result of baroclinic instability. We expect the mechanisms discussed in the present study to be relevant for synoptic eddy feedbacks involved in real atmospheric stratosphere-troposphere coupling, while acknowledging that planetary waves, which are not explicitly considered here, may also play an important role.

More details on how the basic state is constructed are given in the Appendix. Starting from the described initial conditions the experiments are then either run freely (without any external forcing) or including a linear Rayleigh surface friction, following the friction profile specified by Held and Suarez (1994) with a maximum friction coefficient of  $k_f = 1 \text{ day}^{-1}$  at the surface, gradually reducing to zero at 700 hPa ( $z \approx 3 \text{ km}$ ).

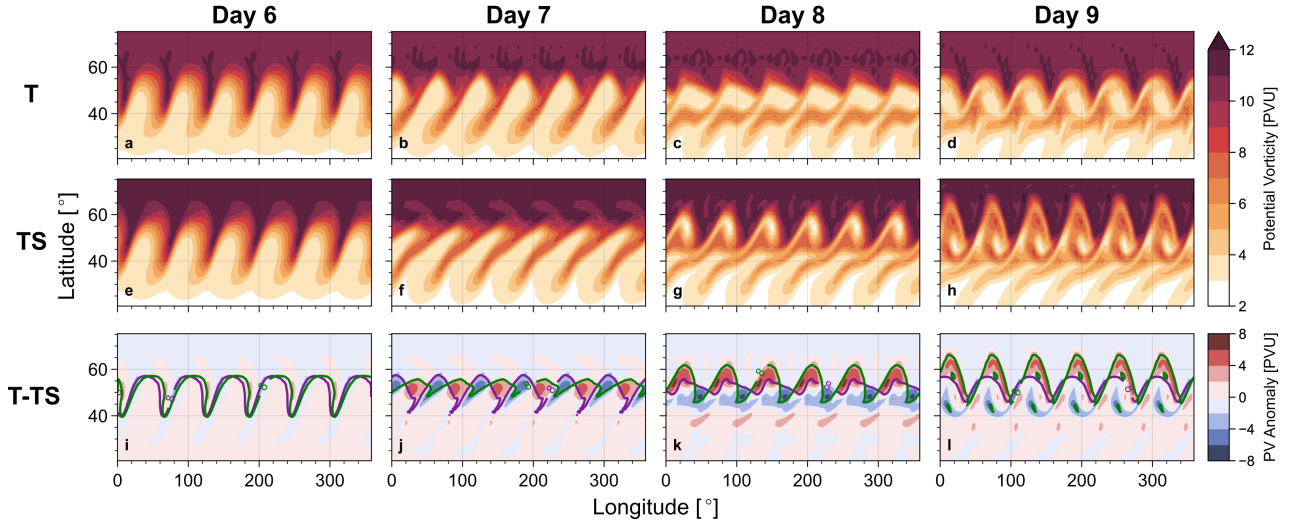
### 3 Modification of the life cycle by a stratospheric jet

We start our study by investigating in what way the general evolution of an idealised baroclinic life cycle is altered when the initial conditions include a tropospheric and a stratospheric jet, the latter representing the winter time polar vortex, compared to when they include a tropospheric jet only, as is usually the case after a SSW and is the conventional life cycle setup. In the rest of this section we therefore analyse a set of life cycle experiments with varying values of the stratospheric jet strength parameter  $u_{Smax}$  (see Equation A2 in the Appendix) and thus varying strength of the stratospheric jet that is added onto the system with tropospheric jet only.

#### 3.1 Modification of the baroclinic wave breaking

The evolution of idealised baroclinic life cycles is often described in terms of the distribution of potential vorticity (PV) on an isentropic surface close to the jet core (or equivalently close to the tropopause). Zonal modulations in PV contours in this region of sharp PV gradient (also seen in Figure 1) give insights into the growth and decay of the eddy field, while any change in the position of the maximum in zonally averaged PV gradient represents a meridional shift of the jet. The top and middle rows in Figure 2 show the horizontal PV distribution on the 350 K isentrope at selected days for the two initial state configurations with tropospheric jet only (experiment T) and tropospheric and stratospheric jet (experiment TS).

200 The general evolution of both experiments is similar to each other in the sense that the baroclinic wave grows gradually until about day 6. At that point the wave becomes non-linear, breaks and eventually decays. ~~However, especially the~~  
Note that the PV distributions at day 6 are almost identical for the experiments T and TS, suggesting that the difference  
in evolution during the linear growth phase induced by stratospheric winds are rather small. This represents an important  
distinction from previous baroclinic life cycle studies that have highlighted stratospheric impacts during the growth phase  
205 (e.g., Wittman et al. (2007); Kunz et al. (2009); Smy and Scott (2009)). We will discuss this apparent contradiction in more  
detail later in this section. In contrast to the small differences during the linear phase, the non-linear decay phase shows substantial differences in the specific evolution of the PV field when a stratospheric jet is present. The wave breaking is still characterised by filaments of high PV that stretch out on the equatorward side of the jet core, break off and eventually roll up anticyclonically, but the timing of events and the details of the small scale structures are altered considerably compared to the  
210 tropospheric jet only case. The decay of the baroclinic wave happens faster and at day 9 a new wave structure seems to have grown already, showing strong characteristics of cyclonic wave breaking (sometimes referred to as LC2 life cycle in contrast to the anticyclonic LC1 life cycle; see, for example, Thorncroft et al. (1993) for further details).



**Figure 2.** Evolution of PV distribution on the 350 K isentrope at different days for a system with tropospheric jet only (experiment T, top panel) or a tropospheric and stratospheric jet (experiment TS, middle panel). The bottom panel shows the difference of both experiments (T-TS), with 8 PVU contours of the respective full fields superimposed.

To highlight the modification in PV evolution induced by the presence of a stratospheric jet the bottom panel of Figure 2 shows the difference in the PV field of a simulation with and without stratospheric jet. Overlaid are the corresponding 8 PVU  
215 contours of the two respective experiments. It can be seen that at day 6, i.e., at the end of the linear growth phase, the two baroclinic waves have a similar magnitude and structure, but are slightly phase shifted with respect to each other. This shift can potentially be explained by a minor increase in phase speed in the case with a stratospheric jet. This might ~~either~~ be due to a minor increase in wind speed near the tropopause (also further discussed in Sections 4 and 5) and/or a change of the

corresponding PV gradient in that region due to the slightly modified wind structure in the case with stratospheric jet. While a  
220 pure zonal phase shift of the wave should not have any influence on the subsequent behaviour of the wave-breaking due to the  
zonal symmetry of the system, it does indicate a change in the dispersion relation.

At days 7 to 9, i.e., during the non-linear phase, the evolution of the system is strongly influenced by the stratospheric jet  
and Figure 2 shows a large difference in PV distribution. Especially at days 8 and 9 the baroclinic wave in experiment TS,  
including a stratospheric jet, seems to have entered a second growth phase, while the wave in experiment T still seems to be  
225 decaying. As mentioned in Section 1 these ~~‘secondary life cycles’~~ during the non-linear decay phase have been discussed  
previously by Barnes and Young (1992). We find the details of the non-linear phase, like the occurrence, timing or apparent  
flavour (in a LC1/LC2 sense) of ‘secondary cycles’, to be very sensitive to small changes of the initial conditions or the details  
of the physical processes involved, as can also be seen in Figure 2. Recall that, as mentioned in Section 2, the occurrence and  
strength of these secondary cycles varied in a set of sensitivity experiments with lower vertical resolution. For the purpose of  
230 this study we therefore focus primarily on the evolution of the entire life cycle, e.g., in terms of the difference between initial  
and some final state.

### 3.2 Dependency on stratospheric jet magnitude

In addition to the evolution of the PV field baroclinic life cycles can be quantified in terms of the global energetics of the system,  
typically with a strong focus on eddy kinetic energy (EKE), which describes the growth and decay of the baroclinic wave in  
235 the region of large meridional PV gradient near the jet core (see Figure 1). In particular the decay of EKE is associated with  
an energy transfer to the zonal mean state, i.e., an increase of the mean kinetic energy (MKE). This increase in MKE can be  
associated with a poleward shift, and a corresponding acceleration, of the tropospheric jet due to wave-mean-flow interactions  
and poleward eddy momentum fluxes during the decay phase of the life cycle.

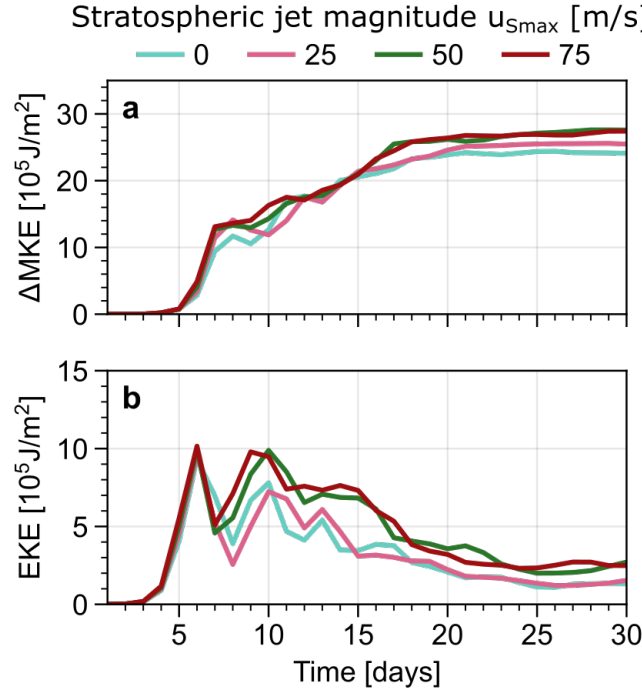
The way the evolution of the life cycle is altered by a stratospheric jet can be seen in terms of EKE and MKE time series,  
240 shown in Figure 3 for experiments with different values for the stratospheric jet magnitude  $u_{Smax}$  (see Appendix for details).  
Note that here we use  $\Delta MKE$ , which is simply the change in MKE with respect to the initial conditions and that both, EKE and  
 $\Delta MKE$ , are displayed as vertically integrated ~~and horizontally~~ (over the entire atmosphere) and horizontally averaged (over  
the northern hemisphere) ~~averaged~~-energy densities.

In agreement with Figure 2, which suggests only a phase shift in the baroclinic waves during the linear phase, but no  
245 difference in magnitudes, Figure 3 shows essentially no sensitivity to introducing a stratospheric jet before day 6, in particular  
we do not find any significant change in growth rate, as has been reported by other authors, e.g., Wittman et al. (2007). A  
potential explanation for the strong change in growth rate found by Wittman et al. (2007) could be a substantial difference in  
meridional PV gradient (due to the substantial modification of the vertical curvature of zonal wind at the tropopause) between  
their different experimental setups. The basic states used in the present study, on the other hand, do only slightly differ in terms  
250 of their tropopause level PV gradients (see Figure 1). However, during the ~~non-linear-phase~~ non-linear phase, so from day 7  
onwards, the stratospheric jet seems to extensively alter the evolution of the life cycle. Especially the onset of a secondary  
phase of wave growth (with EKE peaking again at about day 10) seems to happen about a day earlier when a stratospheric jet is



present in the system, and leads to a much stronger and more persistent secondary peak. The persistently elevated EKE of the secondary cycles during the non-linear phase (with EKE reducing again towards the final state) is consistent with the idea of a stronger LC2 flavour (which is often characterised by persistently increased EKE in the decay phase) of the secondary cycles, as is also suggested by Figure 2 and is further discussed in Section 5.

The alteration of the system as we increase  $u_{Smax}$  does not only manifest as changes in the details of how the wave breaking evolves, but also leads to a change of the final state (here defined as average over days 20-30), in particular a systematic increase of  $\Delta MKE$ .

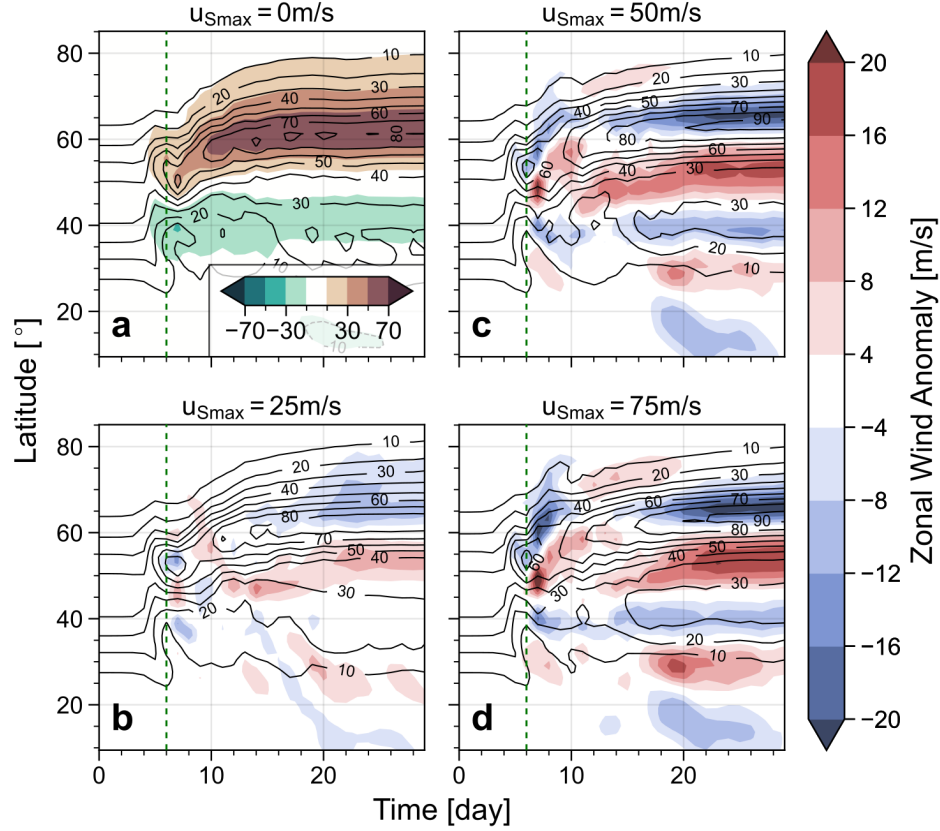


**Figure 3.** Evolution of mean kinetic energy change (a) and eddy kinetic energy (b) of the system with a tropospheric jet and a stratospheric jet with varying strength parameter  $u_{Smax}$  (see Equation A2). The case with  $u_{Smax} = 0$  corresponds to experiment T, the case  $u_{Smax} = 75$  m/s to experiment TS. Energies are displayed as vertically integrated and horizontally averaged energy densities.

The elevated values of EKE during the decay phase and  $\Delta MKE$  in the final state are are consistent with an enhanced barotropic conversion of energy from EKE to MKE (see Fig. S2 in the supplement), as also further discussed in Section 5. The increase in final state  $\Delta MKE$  can further be linked to a stronger poleward shift (and correspondingly a stronger acceleration) of the tropospheric jet during over the course of the life cycle when a stratospheric jet is present.

This relative shift (compared to the experiment T, with tropospheric jet only) can be seen in Figure 4, which shows in all subplots as black contours the evolution of the zonal mean zonal wind field at 10 km. Figure 4a furthermore shows the zonal wind anomaly of experiment T with respect to the initial conditions. One can clearly see a dipole pattern developing around

the initial jet core (45° latitude) at the start of the non-linear phase at about day 6 and strengthening roughly until day 15, corresponding to a poleward shift of the jet core to about 60° latitude [during the life cycle](#).



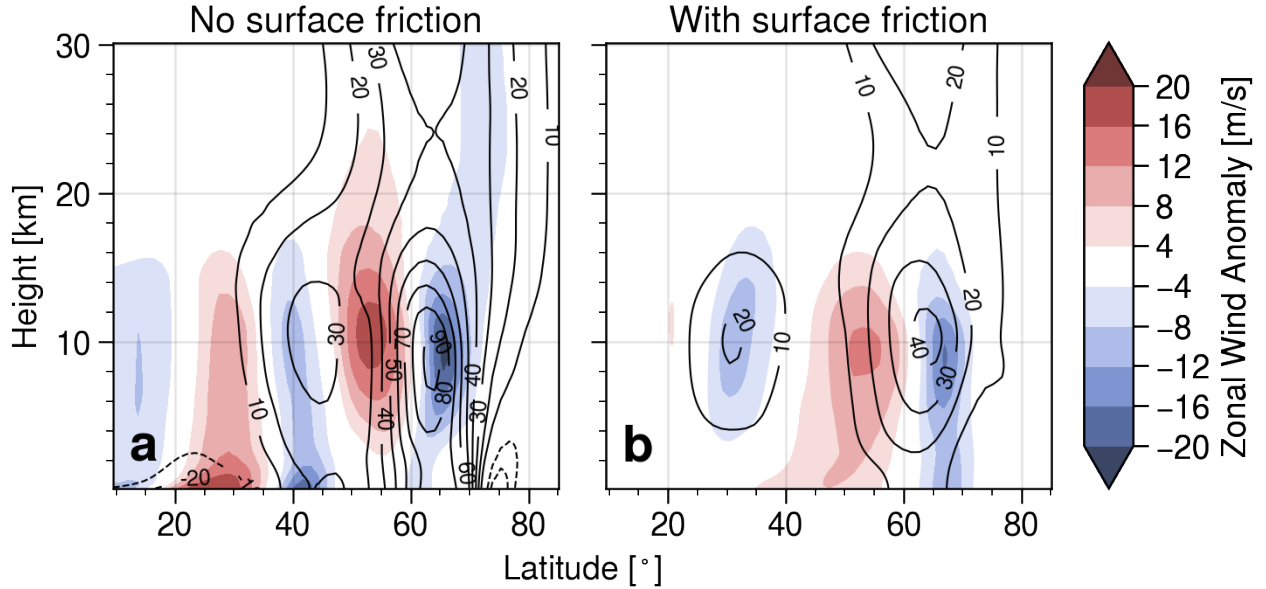
**Figure 4.** Black contours: Evolution of zonal mean zonal wind [m/s] on the 10 km surface for experiments with tropospheric jet only (a) or tropospheric and stratospheric jet of varying strength (b-d); the case with  $u_{Smax} = 0$  corresponds to experiment T, the case  $u_{Smax} = 75$  m/s to experiment TS. The shading in (a) shows the wind anomaly with respect to the initial state, the shading in (b-d) shows the wind anomaly induced when the stratospheric jet is removed from the system [\(i.e., the wind anomalies in b-d are formed by subtracting the respective wind fields from the wind field shown in subplot a\).](#) The vertical dashed lines indicate day 6.

Figures 4b-d show the evolution of zonal mean zonal wind anomaly at 10 km of experiment TS, with varying strength of the imposed stratospheric jet, relative to experiment T, with tropospheric jet only [\(i.e., the difference between the wind field of experiment T and the wind field of the respective experiment\).](#) The displayed anomaly therefore indicates the changes in zonal mean zonal wind induced by a removal of the stratospheric jet from a system (hence modelling changes induced by a SSW). As suggested by the MKE time series shown earlier the zonal wind anomaly evolution indicates a dipole around the position of the final jet core emerging during the non-linear phase of the life cycle. The change in zonal wind corresponds to a stronger poleward shift of the jet during the final state in cases where a stratospheric jet is present, or equivalently, a relative equatorward

shift of the tropospheric jet when the stratospheric jet is removed. This jet shift is analogous to the NAM-like signature that has been observed after SSW events. It further indicates the importance of tropospheric synoptic-scale eddy feedback in causing the observed negative NAM-signal, as has previously been shown by other studies (e.g., Domeisen et al., 2013; Hitchcock and Simpson, 2014), and allows for a simple way to quantify this eddy feedback.

### 280 3.3 Vertical structure of the response and influence of surface friction

The vertical structure of the relative jet shift of the final state can be seen in Figure 5, showing the difference in zonal mean zonal wind during the final state (days 20-30 mean) between experiments T (with tropospheric jet only) and TS (also including a stratospheric jet of magnitude  $u_{Smax} = 75$  m/s). Subplot 5a shows the latitude-height equivalent of subplot 4d averaged over the final state, while subplot 5b illustrates the corresponding zonal wind anomaly for an experiment with surface friction applied to the system (see Section 2 for details). Both subplots show a clear equatorward jet-shift signature around the jet core of the final jet when the stratospheric jet is removed. Note that the inclusion of surface friction will lead to a continuous dissipation of energy and thus a steady ‘final state’ is not reached (see supplementary material). However, we define the final state analogous to the case without friction (days 20-30 average) in order to analyse the effect of surface friction on the evolution of the main life cycle.



**Figure 5.** Contours: zonal mean zonal wind [m/s] of the final state (days 20-30 average) of a system with stratospheric jet. Shading: Changes to the final state zonal mean zonal wind when the stratospheric jet is removed from the system. Subplot a) shows an experiment without surface friction, while b) displays an experiment with surface friction as described in Section 2.

290 ~~However, several~~ Several differences can be observed in the final state of the life cycle for systems with and without surface friction. First, an overall weakening of the jet in the final state (black contours) can be observed when surface friction is included, which is easily explained by the direct dissipation of kinetic energy over the course of the life cycle due to the added friction process. The same argument holds for the disappearance of the strong wind anomaly patterns close to the surface at about 30° and 40° latitude in the case without friction. These patterns develop due to strong temperature fluxes in this region  
 295 arising from the large meridional surface temperature gradient (see Figure 1) and they are likely not influential on the standard baroclinic life cycle evolution. More importantly, however, the vertical structure of the dipole pattern around the final jet core at 60° latitude is drastically different between the experiments with and without surface friction displayed in Figure 5. When the system is subject to surface friction during the life cycle the corresponding dipole pattern is more barotropic, thus it extends much further down and shows much stronger anomalies at the surface.

300 Figures 4 and 5 indicate a tendency of the tropospheric jet to exhibit a weaker poleward shift during the baroclinic life cycle if there is no stratospheric jet present compared to when there is. This behaviour is consistent with the negative NAM response, associated with an equatorward shift of the tropospheric jet, observed during periods following SSWs (see Baldwin and Dunkerton, 2001). It further provides a simple way to quantify the eddy feedback processes potentially involved in creating the corresponding jet shift signal. Figure 5 shows the shift signal only to have a significant surface contribution if the system is  
 305 subject to surface friction.

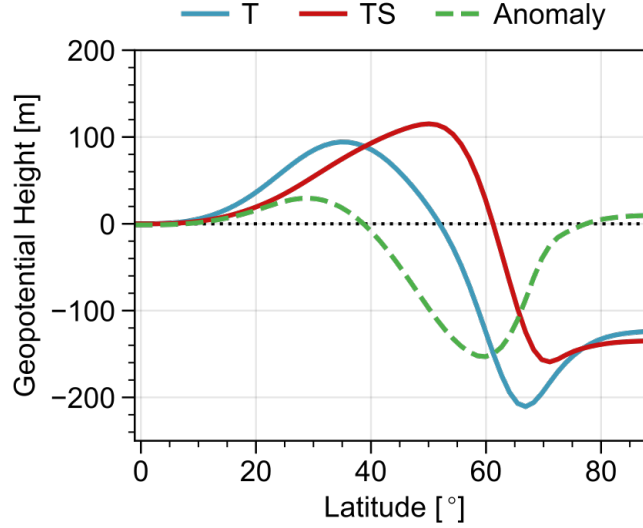
To further illustrate the surface signal observed in our model experiments Figure 6 shows the geostrophic geopotential height field  $Z$ , calculated by solving the equation

$$\partial_\phi Z = -fa\bar{u} - \bar{u}^2 \tan \phi \quad (2)$$

via simple numerical integration with boundary condition  $Z(\phi = 0) = 0$  for the zonal mean zonal wind field  $\bar{u}$  of the fi-  
 310 nal state (note that we neglected the surface friction term in Equation 2 as it tends to be small during the final state of our experiments). Here  $f$  is the Coriolis parameter,  $a$  the radius of the Earth,  $g$  the gravitational acceleration and  $\phi$  the latitude. Since  $\bar{u}(z = 0)$  vanishes for the initial state the surface geopotential height  $Z_{sfc} \equiv Z(z = 0)$  of the final state (or more precisely its gradient) describes the change in surface winds induced over the course of the baroclinic life cycle.

Figure 6 shows  $Z_{sfc}$  for experiments that include surface friction and two different sets of initial conditions: T and TS, i.e.,  
 315 including a tropospheric jet only and including both, a tropospheric and a stratospheric jet. For both experiments we find the development of strong meridional gradients in  $Z_{sfc}$  at around 50° or 60° latitude, respectively, consistent with strong surface winds. The farther equatorwards shifted position of the gradient of  $Z_{sfc}$  in experiment T, relative to experiment TS, indicates again the relative equatorward shift of the final tropospheric jet if the stratospheric jet is removed from the initial conditions, corresponding to the NAM-signal discussed earlier.

320 The strength of the NAM-like jet shift signal depends on the magnitude of the stratospheric jet ( $u_{Smax}$ ) included in the system, as can be seen in Figure 7. First, the NAM-signal, in the form of a dipole jet shift pattern around 60° latitude, seems to develop for stratospheric jet magnitudes below about  $u_{Smax} \lesssim 50$  m/s, but stays mostly unchanged for stratospheric jets



**Figure 6.** Zonal mean geopotential height at 1000hPa (or equivalently  $z = 0$ ) of the final state for two experiments with surface friction and with tropospheric jet only (T) and tropospheric and stratospheric jet (TS), respectively. The dashed line shows the difference of both.

exceeding  $u_{Smax} \gtrsim 50$  m/s. Second, the NAM-response does not seem to be symmetric for positive and negative values of  $u_{Smax}$ . While the jet shift signal develops already for relatively weak westerly stratospheric jets, no coherent ~~such~~ signal can  
 325 be observed for easterly stratospheric jets for the parameter range shown (a positive NAM-signal, i.e., a relative poleward shift, only starts to develop for  $u_{Smax} \lesssim -40$  m/s).

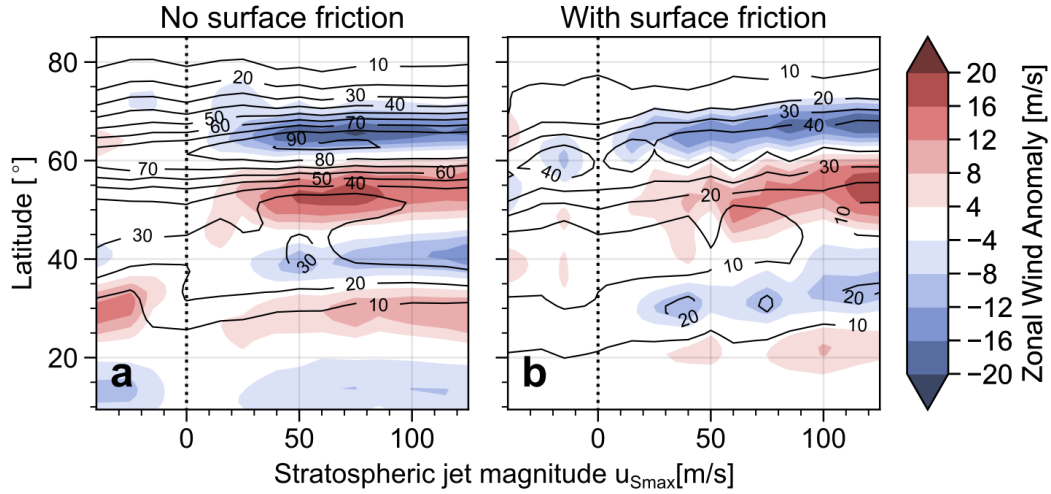
In the rest of this paper we investigate the influence of a stratospheric jet onto the final state of the baroclinic life cycle and the resulting NAM-like signature in more detail. In particular we identify a region in the lower stratosphere which is highly sensitive to changes in the zonal wind that are induced by the inclusion of a stratospheric jet.

#### 330 4 Sensitivity of the life cycle to changes in the extratropical lower stratosphere

In the previous section we established that introducing a stratospheric jet can modify the evolution of the system in an idealised baroclinic life experiment, as has also been shown by other authors (e.g., Wittman et al., 2004). In this section we show that the system is particularly sensitive to changes in wind structure in the extratropical lower stratosphere (heights below about 25 km), while changes in the middle and upper stratosphere have almost no influence on the final state. In order to investigate this  
 335 sensitivity we analyse a set of experiments with initial conditions that include a tropospheric jet, as well as a stratospheric jet with modified vertical structure.

We modify the structure by multiplying the profile of the stratospheric jet used in experiment TS by a function  $\eta(z)$  (see Equation A2). We choose  $\eta(z)$  to follow a tanh-profile, which allows us to smoothly set the winds of the stratospheric jet



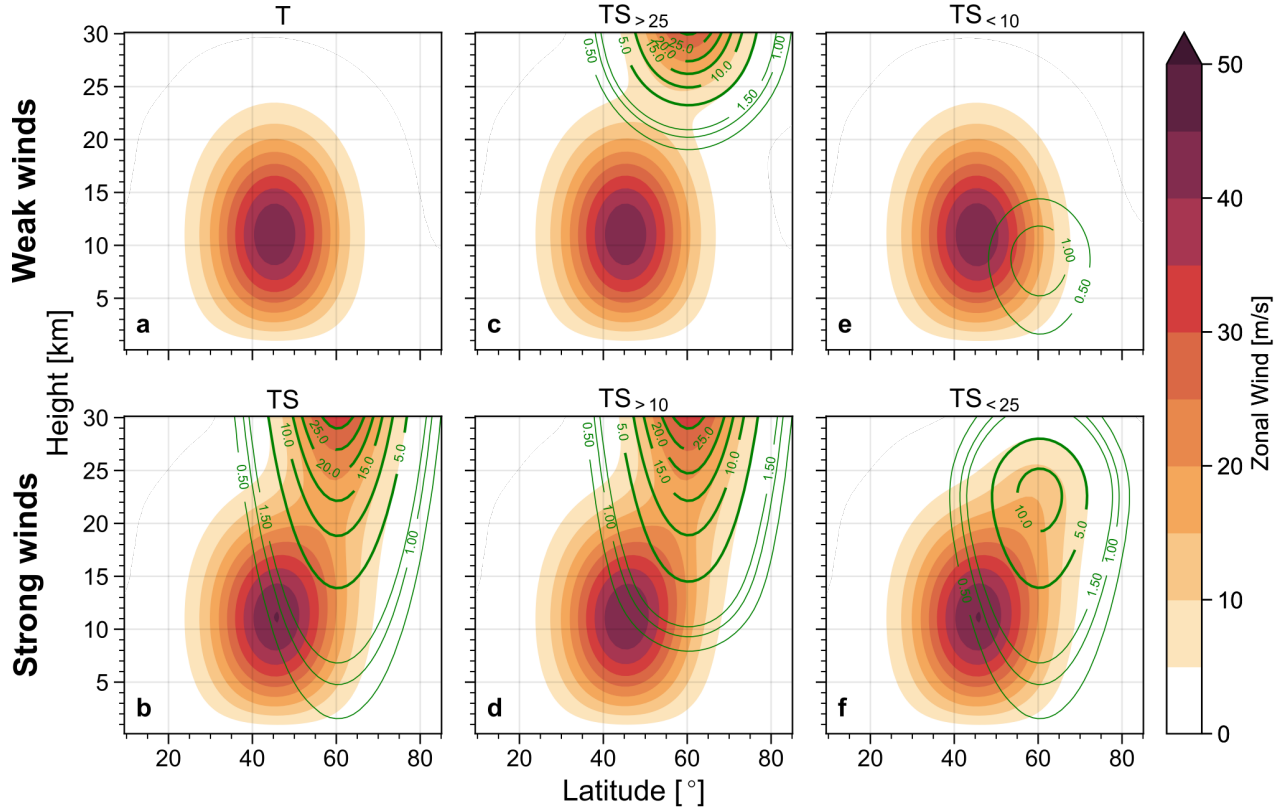


**Figure 7.** Contours: Zonal mean zonal wind at 10 km of the final state of experiments that include a stratospheric jet with varying strength parameter  $u_{Smax}$ . Shading: The changes induced when the stratospheric jet is removed from the system. The two subplots show experiment without and with surface friction, respectively. The vertical dotted line indicates  $u_{Smax} = 0$ , and thus experiment T.

component to zero below or above a set transition height  $z_\eta$  and thus investigate which part of the stratospheric jet has the  
 340 strongest influence on the life cycle. We hereby refer to the experiments where we only include the part of the stratospheric  
 jet below height  $z_\eta$  as  $\text{TS}_{<z_\eta}$ , and correspondingly refer to the experiments where we keep the part above  $z_\eta$  as  $\text{TS}_{>z_\eta}$ ,  
 (for simplicity we drop the units of  $z_\eta$  within this notation and set it to be kilometres). See the Appendix for details on how the  
 basic state is defined.

Figure 8 illustrates the different basic states in terms of the full zonal mean zonal wind field, and the anomaly with respect  
 345 to experiment T, i.e., the experiment without any superimposed stratospheric jet. Subplots 8a and b show experiments T and  
 TS, including no or the full stratospheric jet, respectively. The experiments displayed in subplots 8b and c only superimpose  
 the upper part of the stratospheric jet, above either 25 km or 10 km, while the experiments displayed in subplots 8e and f only  
 include the respective lower parts.

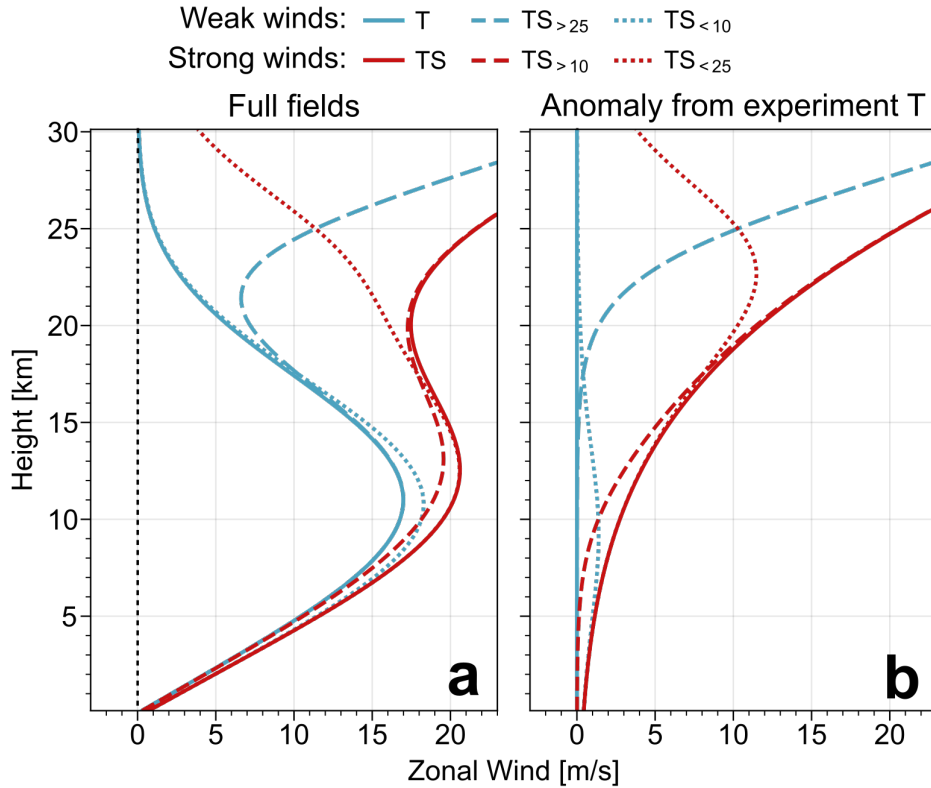
Details of the vertical structure of the various initial wind fields can also be seen in Figure 9, displaying the zonal wind at  
 350  $60^\circ$  latitude, i.e., at the northern flank of the tropospheric jet and through the core of the stratospheric jet. A very prominent  
 difference is that profiles where the stratospheric jet reaches into the lower stratosphere have substantially increased wind  
 speeds in that region (roughly between 10 and 25km), compared to profiles where the contribution of the jet is mostly confined  
 to the troposphere or the middle and upper stratosphere. This criterion divides the six profiles into two groups, 'Set-1'-one  
 consisting of profiles T,  $\text{TS}_{>25}$  and  $\text{TS}_{<10}$  with weak winds in the lower stratosphere, and 'Set-2'-the other consisting of  
 355 profiles TS,  $\text{TS}_{>10}$  and  $\text{TS}_{<25}$  with strong winds in the lower stratosphere. In most of the rest of this section we analyse the  
 experiments with different initial conditions keeping in mind the grouping into these two sets.



**Figure 8.** Zonal wind (shading) and anomaly from experiment T (green contours, in m/s) of the initial conditions for experiments with a tropospheric jet and varying vertical profiles of the superimposed stratospheric jet depending on the function  $\eta(z)$  in Equation A2. Note that both sets of green contours, thick and thin, show the same quantity, but for different level ranges.

To visualise the NAM-like jet shift signature of the final state, and to investigate which contribution to this jet shift can be associated the different parts of the stratospheric jet, Figure 10 shows the zonal mean zonal wind averaged over days 20-30 and the corresponding anomaly from experiment T (with tropospheric jet only).

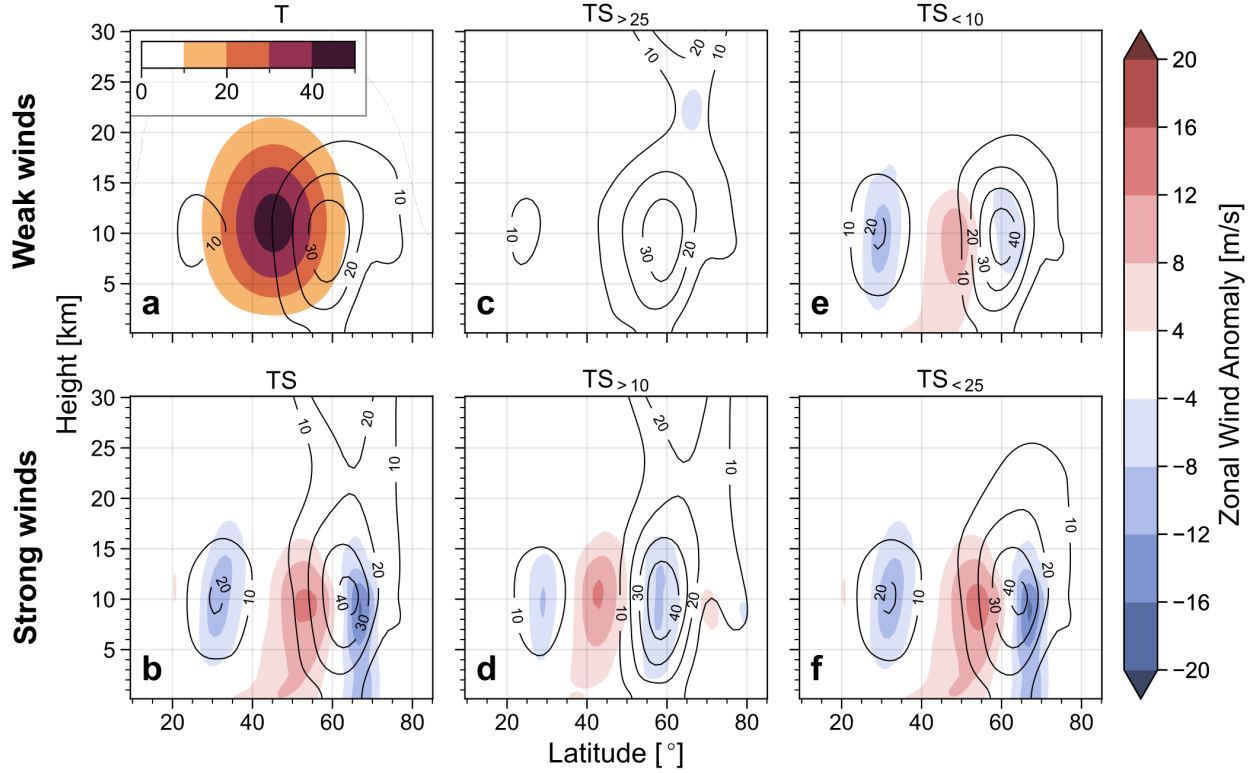
360 We first look at the experiments ~~of Set 1, with weak winds in the lower stratosphere.~~ The final state zonal wind field of experiment  $TS_{>25}$  (Figure 10c) does not show any substantial deviation from experiment T, indicating that winds in the middle and upper stratosphere have virtually no influence on the life cycle. Experiment  $TS_{<10}$ , with superimposed winds confined to the troposphere, shows a dipole pattern, which could potentially be attributed to the projection of the wind modification on e.g., the increase in tropospheric jet magnitude or the vertical shear, also further ~~discusses~~ discussed in Section 5. However, also  
365 note that the superimposed winds of the stratospheric jet do not abruptly vanish at the given cut-off height (e.g., above 10 km for  $TS_{<10}$ ), but follow a smooth transition over the course of about 4 km and therefore still reach into the lower stratosphere region.



**Figure 9.** Vertical profiles of zonal mean zonal wind at  $60^\circ$  latitude of the initial conditions for different experiments without (T), with full (TS) or with partial stratospheric jet (other profiles). Subplot a) shows the full fields, subplot b) the anomaly from experiment T.

The experiments ~~of Set 2~~ with strong winds in the lower stratosphere (bottom panel of Figure 10) do all show a clear dipole structure in the anomaly field, centred around about  $60^\circ$  latitude. Note in particular the strong signal of experiment   
 370  $TS_{<25}$ , where the superimposed winds are confined to the troposphere and lower stratosphere, further suggesting the winds in the middle and upper troposphere ~~to~~ have no significant contribution in causing the observed jet shift. Experiment  $TS_{>10}$ , including a stratospheric jet that reaches into the lower stratosphere but does ~~reach not not reach~~ far into the troposphere, also shows a clear dipole pattern in zonal wind anomaly. In particular compare experiments  $TS_{>10}$  and  $TS_{>25}$ , as well as  $TS_{<25}$  and  $TS_{<10}$ : in both cases ~~does~~ the jet shift signal increase-increases in strength when the superimposed stratospheric jet reaches   
 375 into the lower stratosphere (10 km to 25 km), compared to when it does not. The significance of the lower stratospheric wind anomalies ~~are-is~~ discussed further in Section 5.

The surface signal of the NAM-like response discussed above can be seen in Figure 11, displaying the zonally averaged ~~geostrophic~~ geopotential height field calculated via Equation 2. It can clearly be seen how the different experiments show indications for NAM-like surface signals in good agreement with what is shown in Figure 10. Especially the experiments ~~of~~



**Figure 10.** Black contours show the zonal mean zonal wind [m/s] distribution of the final state for different experiments. The shading in (a) indicates the initial zonal winds [m/s] of experiment T, the shading in (b-f) shows the anomaly from experiment T in zonal mean zonal wind of the final state. All experiments include surface friction.

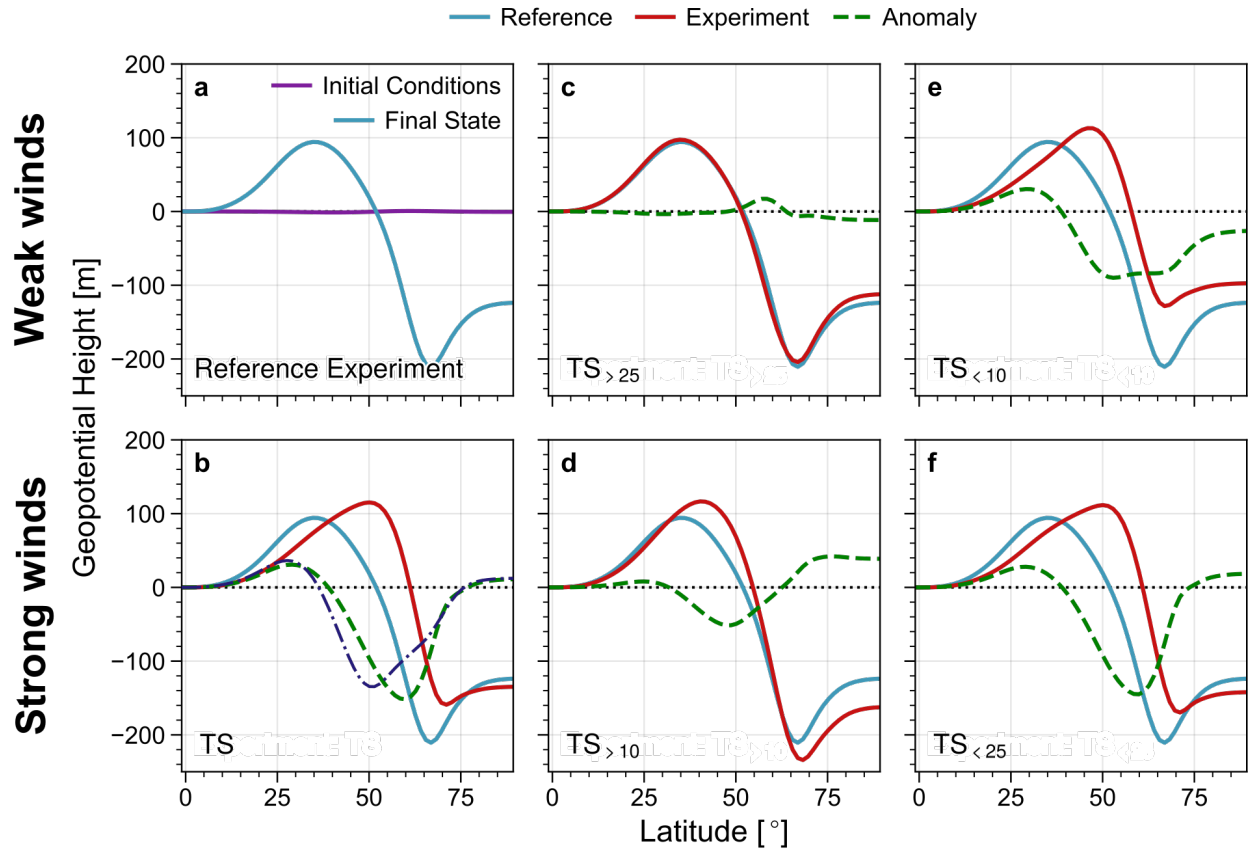
380 Set-2-with strong lower-stratospheric winds (bottom panel) show a poleward shift and acceleration of the surface winds (in terms of gradient of the shown curves) relative to the reference experiment T, with only the tropospheric jet.

Figure 11b further shows the sum of the geopotential height anomalies induced by removal of the (partial) stratospheric jet from the experiments  $T_{<10}$  and  $T_{>10}$ , i.e, experiments where we only include the part of the stratospheric jet above or below 10km. The similarity of this sum to the corresponding geopotential height anomaly of experiment TS, with full stratospheric jet included, suggests a certain additivity of the response to the stratospheric jet<sup>1</sup>, also further discussed in Section 5.

## 5 Discussion

In Section 3 we showed that the growth phase of an idealised baroclinic life cycle is almost unchanged when introducing a stratospheric jet to the system. In particular the linear growth rate of the baroclinic wave (in terms of EKE) is not sensitive to

<sup>1</sup>Note that the same similarity seems to hold for the sum of anomalies of experiments  $T_{<25}$  and  $T_{>25}$  (not shown explicitly).



**Figure 11.** Geopotential height at 1000 hPa of the final state of the experiments displayed in Figure 8. Subplot a) illustrates the difference in initial and final state geopotential height for experiment T, all other subplots illustrate the difference between the respective experiment and the reference experiment T. The purple dash-dotted line in subplot b) shows the sum of the anomalies for the cases  $T_{<10}$  and  $T_{>10}$ . All experiments include surface friction.

changes in stratospheric conditions. These findings seem to contradict previously reported results (e.g.,  
 390 [Wittman et al. \(2007\)](#); [Kunz et al. \(2009\)](#); [Smy and Scott \(2009\)](#)). A likely explanation for this discrepancy is that the linear growth phase is highly sensitive to tropopause-level PV gradients of the initial conditions, which had been altered due to changes in the stratosphere in these previous studies. The structure of the PV gradient in our experiments, on the other hand, is essentially not altered by the inclusion of stratospheric winds and the linear growth of baroclinic waves (driven by tropospheric heat fluxes) is therefore unchanged. However, we found the presence of a stratospheric jet to substantially alter the non-linear  
 395 decay phase of a baroclinic life cycle. In particular, notably it is during this decay phase where the eddy momentum flux acts to convert eddy kinetic to mean kinetic energy while at the same time driving a meridional shift of the tropospheric jet. Given

that the momentum flux is proportional to the equatorward wave activity flux our findings are consistent with a stratospheric influence via modifications to meridional wave propagation near the tropopause.

Figures 2 and 3 showed changes in the secondary cycles occurring during the non-linear decay stage, including changes in number, strength, duration, timing and apparent type (or flavour) of these secondary cycles. The different types of baroclinic wave breaking (LC1 and LC2) have been linked to different weather regimes, and thus corresponding transitions within a life cycle can potentially have a large impact on surface weather (e.g., Michel and Rivière, 2011). As discussed, the baroclinic decay phase of experiment TS shows characteristics of both, LC1 and LC2 flavour, or equivalently cyclonic and anticyclonic wave breaking, while experiment T shows more LC1 characteristics. However, the general behaviour of the TS life cycle (e.g., in terms of its final state) still follows primarily the (anticyclonic) LC1 type, and it only seems to experience individual, transient (cyclonic) LC2 wave-breaking events. The importance of these transient LC2 events for the overall meteorological regime is presently not clear. Note that other authors have previously also reported about transitions between LC1 and LC2 wave breaking states based on stratospheric conditions (e.g., Kunz et al., 2009), but mostly in terms of the entire life cycle, rather than in terms of more transient events.

~~The~~

Further note that the introduction of a stratospheric jet not only modifies the details of the wave breaking during the decay phase, but also the (~~steady~~quasi-steady) final state of the life cycle. In particular, do generally elevated values of EKE during the decay phase and increased MKE during the final state suggest an enhanced barotropic EKE to MKE conversion (as also seen in Fig. S2 in the supplementary material). The inclusion of a stratospheric jet to a basic state also formally  
corresponds to an increase in MPE (mean potential energy, also referred to as available potential energy or APE), where the latter in principle forms the main energy source for the life cycle and is, via EKE, ultimately converted into MKE. However, since this increase in MPE is primarily associated with stratospheric temperature structure, it is unlikely to contribute to the predominantly tropospheric energy conversions during the life cycle.

Consistent with the change in final state MKE we observed a relative equatorward shift of the tropospheric jet in the final state when removing the stratospheric jet from the initial conditions of the system, as can be seen in Figures 5 and 6. This relative jet shift is analogous to the NAM-like signature that has been observed after SSW events. To what extent the observed NAM response to SSWs is similarly influenced by tropospheric eddy feedbacks, as suggested by our results, remains to be quantified further.

~~It is worth to~~ It also is important to remember that the coupling of troposphere and stratosphere works in both directions and the coupled system will generally react as a whole to any (tropospheric or stratospheric) forcing.

We further want to point out that the relative meridional shift between the final jet in the experiments T and TS results from differences in the meridional eddy momentum transport during the life cycle (not shown). The increased momentum fluxes in experiment TS, compared to T, can be related to increased wave activity around tropopause level, which is consistent with the increase in EKE shown in Figure 3. Similar changes in eddy momentum transport as a result to (in particular lower-stratospheric) climate anomalies have been observed previously in idealised general circulation model experiments by various authors (e.g., Polvani and Kushner, 2002; Butler et al., 2010).



Figure 5 further shows that the surface signal of the NAM-response in the final state, given by the zonal mean zonal wind difference between experiments T and TS, is enhanced when the system is subject to surface friction. The effect of surface friction to increase the surface wind signal of the NAM-response might seem counter-intuitive. However, as we already pointed out in  
435 Subsection 1.3, surface friction can provide a way for tropopause level eddy momentum fluxes to couple to the surface winds. The modification of the baroclinic eddy field by the presence of a stratospheric jet can therefore project more strongly onto the surface winds and produce a stronger surface signal. The evolution equation of the vertically averaged zonal mean zonal wind (Equation 1) is often used to argue that on long time scales (where  $\partial_t[\bar{u}] \approx 0$ ) the eddy flux convergence has to be balanced by the (dissipation of) surface winds. In our (transient) life cycle experiments we cannot neglect the wind tendency term and  
440 the main balance is given by  $\partial_t[\bar{u}] \approx -\partial_y[\bar{u}'v']$ . However, the dissipation term  $\bar{u}_{sf}/\tau$  provides an important contribution to the equation and strongly modifies the acceleration of the jet, as is already suggested by the factor 2 difference of the final jet magnitude in the cases with and without surface friction (see Figure 5).

When interpreting Equation 1 one also has to keep in mind that it describes the evolution of the full (vertically and zonally averaged) wind field, whereas we are mostly interested in the enhanced surface signal of the difference in wind field between  
445 experiments TS and T (see Figure 5), i.e., the NAM-like shift signal. The line of argumentation, however, is analogous. The introduction (or removal) of a stratospheric jet influences the evolution of baroclinic eddies at tropopause level. Following Equation 1, the corresponding changes in eddy momentum flux then induce changes in the wind tendency (which will tend to be close to the level of the eddy flux, primarily near the tropopause), but also couple directly to the surface winds.

The enhancement of the surface signal by surface friction can potentially be understood via the following mechanism:  
450 the decay stage of the life cycle is characterised by a barotropisation of the tropospheric jet and thus a reduction of vertical shear and a strengthening of surface winds. The surface friction, on the other hand, tends to increase the vertical wind shear and ~~acts-therefore act~~ as a source of baroclinicity. This increase in baroclinicity then leads to ~~additional-(secondary)-life~~  
~~eyeles-an enhanced barotropisation of the jet (similar to what was observed by Barnes and Young, 1992) and a correspondingly~~  
~~enhanced downward propagation of the jet-shift signal. The idea of an enhanced baroclinisation/barotropisation is consistent~~  
455 ~~with the formation of additional life cycles which we observed~~ during the late stages of the life cycle ~~(seen-around days 16-25~~  
~~in-Figure-S2-in experiments with surface friction (seen roughly between days 15 and 25 in Fig. S3 in the supplementary~~  
~~material)and thus an additional barotropisation of the jet, enhancing the downward propagation of the jet-shift signal.~~

Figure 7 indicated that the strength of the NAM-response following the removal of the stratospheric jet depends non-linearly on the magnitude of the stratospheric jet. In particular, the signal seems to saturate when the stratospheric jet magnitude exceeds  
460 a certain value and stronger jets do not lead to a stronger NAM-signal any more. This behaviour suggests that an anomalously strong polar vortex does not necessarily lead to anomalously positive NAM-signals. Similarly Figure 7 indicates that a reversal of the stratospheric jet (with  $u_{Smax} < 0$ ) does not lead to a negative NAM-response with respect to experiment T, which suggests that in terms of NAM-response it is not important ~~weather-whether~~ a SSW leads to slightly or strongly reversed winds of the polar vortex. However, the setup of baroclinic life cycle experiments does, of course, not capture the dynamics around  
465 SSWs in their entire complexity and these results do not necessarily carry over to the real atmosphere.

In Section 4 we showed that the NAM-response observed in the final state of our life cycle experiments is mostly caused by the change in wind structure in the lower stratosphere when including the stratospheric jet, rather than wind anomalies in the middle and upper stratosphere (see Figure 10). ~~However, it~~ The sensitivity of the eddy feedback to wind anomalies in the lower stratosphere is consistent with results previously reported by various authors (e.g., Butler et al., 2010).

470 It should be noted that changing the wind structure in the lower stratosphere does also introduce changes in various other characteristics of the corresponding initial conditions, like the height of maximum wind speed, the vertical wind shear in the upper troposphere (roughly up to 10km) and the magnitude of the tropospheric jet (especially obvious for profiles T and TS in Figure 9). However, these three characteristics are intrinsically not completely independent and can potentially all affect the evolution of the life cycle. This can be seen, e.g., since the vertical wind shear is (via thermal wind balance) related to  
475 the horizontal temperature gradient, which drives the growth of baroclinic waves and can, among other things, modify their (linear) growth rate (although note that the near surface shear is almost identical in the different experiments).

We performed a set of sensitivity experiments (not shown) with tropospheric jet only and varying tropospheric jet magnitude (and therefore increased vertical shear in the troposphere). We found that an increase in tropospheric jet strength also leads to a increased poleward shift during the life cycle (i.e., an equatorward shift of the jet in the final state of experiment T relative to  
480 a case with stronger tropospheric jet), similar to the shift observed in Figure 5a. In order to achieve a jet shift signal of similar magnitude as the one shown in Figure 5, however, it was necessary to increase the jet magnitude by order of 10 m/s in these sensitivity experiments (the difference in tropospheric jet magnitude between experiments T and TS is only of the order of 1 m/s.), indicating that other characteristics of the initial state need to contribute and the observed jet shift cannot purely be a result of a strengthened tropospheric jet. The inclusion of the stratospheric jet does to some extend project onto the mentioned  
485 characteristics (e.g., height of the jet core and tropospheric shear) of the total zonal wind profile and the resulting jet shift can potentially be interpreted as the result of a combination of factors.

Figure 11 further suggested that we essentially recover the surface geopotential height signal of experiment TS (with full stratospheric jet), when adding the corresponding signals of experiments  $T_{<10}$  and  $T_{>10}$ . Such additivity of responses might be another indication that the stratospheric jet projects onto various other structures and characteristics (e.g., tropospheric  
490 shear and jet core height) and the corresponding jet shift response forms as the result of a combination of responses to those modifications. However, while the anomalies of the respective experiments seem to be additive when it comes to the surface geopotential height (although not perfect), the middle tropospheric jet shift response in Figure 10 does not appear to follow the same additive behaviour.

As discussed, Figure 10 ~~suggests~~ shows the NAM-like jet shift signature of the life cycle due to the inclusion of a strato-  
495 spheric jet to be mainly caused by the corresponding change in winds in the lower stratosphere, rather than the winds in the middle and upper stratosphere, where the stratospheric jet itself is strongest. A similar conclusion can be drawn from the energetics of the system (provided as supplementary material), which shows a consistent increase in MKE of the final state for the experiments in Set 2 with strong winds in the lower stratosphere (as defined in Section 4), compared to the experiments in Set 1 with weak winds, in a system that does not include surface friction. As also explained in Section 3, this increase in  
500 MKE is caused by the relative meridional shift of the final tropospheric jet. Note that if the system includes surface friction the

constant dissipation of winds leads to a gradual and flow dependent decrease of MKE, which makes the interpretation of the energetics in terms of a final state difficult. However, we find basic states which include a stratospheric jet to be associated with an enhanced barotropic energy conversion of EKE to MKE during the life cycle in both, systems that do and do not include surface friction (see Fig. S2 in the supplement).

505 While the present study discusses the sensitivity of the tropospheric jet shift to the presence of a stratospheric jet during baroclinic life cycles in some detail, various questions remain open and provide potential for future work. It might be possible to use the idealised life cycle set-up discussed above to gain insights into the distinction between the responses to different types of SSWs. For example, sudden stratospheric warming events characterised as split or displacement have been found to be associated with different lower-stratospheric wind anomalies, while the question of differences in their tropospheric  
510 response has not been fully answered (see Charlton and Polvani, 2007; Mitchell et al., 2011; Maycock and Hitchcock, 2015). Smy and Scott (2009) studied the distinction of split and displacement events in idealised life cycle experiments and found strong differences in tropospheric response. However, it should be noted that their experiments are initialised with a basic state constructed via the inversion of an imposed PV field which will inevitably have an influence on the tropospheric initial state and thus directly affect the evolution of the life cycle. The distinct stratospheric influences via direct remote PV signatures or  
515 indirect influences on non-linear baroclinic eddy dynamics remains to be investigated further.

## 6 Summary and conclusions

In this paper we discussed changes in the evolution of idealised baroclinic life cycles induced by the presence of a stratospheric jet. Particular focus was given on a jet shift signal in the zonal wind anomaly of the final state of the life cycle, similar to the signature of negative (surface) anomalies of the northern annular mode (NAM) often observed after sudden stratospheric  
520 warming (SSW) events.

We found that the final state of the life cycle is associated with increased zonal mean kinetic energy when a stratospheric jet is included in the system, roughly representing the polar vortex of typical winter-time conditions, compared to the typical life cycle setup including only a tropospheric jet, roughly representing post-SSW conditions. This increase in mean kinetic energy corresponds to a negative NAM signal in the final state zonal wind, i.e., a relative equatorward shift of the tropospheric jet in  
525 the case with tropospheric jet only compared to the case with tropospheric and stratospheric jet. The negative NAM signal is the result of a reduced poleward shift over the course of the life cycle induced by a reduction in eddy momentum transport at tropopause level.

The corresponding NAM-like jet-shift response has an increased surface signal if the system includes surface ~~fiction~~friction, which might seem counter-intuitive, but is consistent with the idea of an increased coupling of surface winds to the eddy  
530 momentum transport at tropopause level due to the friction.

We further showed that the system is mainly sensitive to changes of the wind structure in the lower stratosphere (heights between 10 km and 25 km), rather than to zonal wind anomalies in the middle and upper stratosphere.

The findings of this paper ~~improve our basic understanding of the weather and climate system in the mid-latitude troposphere and lower stratosphere~~ provide further evidence for the role of tropospheric eddy feedbacks in shaping the tropospheric response to stratospheric events. In particular ~~they provide a potential explanation for the downward propagation of zonal wind anomalies from the stratosphere into the troposphere and the related negative NAM signal observed after~~, they help to explain the observed negative surface NAM signal following SSWs. The ~~idealised life cycle setting further provides a quantitative way to analyse the importance of the tropospheric eddy feedback during and after SSWs~~. ~~simplified nature of the idealised life cycle setup allows for a clean separation of tropospheric eddy feedbacks in the surface response to different stratospheric conditions, highlighting the role of tropopause-level momentum fluxes in the non-linear phase of the life cycle. It furthermore offers quantitative insights into the role of surface friction in modulating the surface response to stratospheric events in a simplified setting.~~

## Appendix A: Appendix: Construction of initial state

The basic state used to initialise our experiments is defined via a zonally symmetric zonal wind field, consisting of two individual components: a tropospheric jet  $U_T$  (representing the mid-latitude jet) and a stratospheric jet  $U_S$  (representing the polar vortex). The total wind field is then given by the sum of both components  $U = U_T + U_S$ , with the tropospheric jet profile being given by

$$U_T = u_{Tmax} (z/z_{Tmid}) \exp((1 - (z/z_{Tmid})^\alpha) / \alpha) \sin^3(\pi \sin^2(\phi)), \quad (A1)$$

where  $z = -H \ln(p/p_0)$  is a log-pressure coordinate with scale height  $H = 7.5$  km and reference pressure  $p_0 = 1000$  hPa and  $\phi$  describes latitude. The parameters  $u_{Tmax}$ ,  $z_{Tmid}$  and  $\alpha$  can be used to modify the jet strength, the core height and the depth of the jet, respectively. The corresponding stratospheric jet profile is defined via

$$U_S = u_{Smax} \eta(z) \exp(-(z - z_{Smid})^2 / \Delta z_S^2 - (\phi - \phi_S)^2 / \Delta \phi_S^2), \quad (A2)$$

where  $u_{Smax}$  determines the strength of the jet,  $z_{Smid}$  and  $\phi_S$  its core position and  $\Delta z_S$  and  $\Delta \phi_S$  its width and depth, respectively. Note that we restrict both jet profiles to the northern hemisphere, i.e., for  $\phi < 0$  we choose  $u_{Tmax} = u_{Smax} = 0$  and therefore keep the southern hemisphere of the basic state at rest.

The function  $\eta(z)$  can be used to further modify the vertical structure of the stratospheric jet. For all experiments in Section 3 we choose  $\eta \equiv 1$ , so the stratospheric jet is unmodified, while for the cut-off experiment in Section 4 we choose

$$\eta(z) = 0.5 (1 \pm \tanh((z - z_\eta) / \Delta z_\eta)), \quad (A3)$$

in order to set the stratospheric jet strength to zero above or below (depending on whether a plus or minus is used within Equation A3) the transition height  $z_\eta$  with a smooth transition of depth  $\Delta z_\eta$ . This gives us a way to isolate the parts of the

560 stratospheric jet within the troposphere, lower stratosphere or middle and upper stratosphere, respectively, and thus study the corresponding influence on the life cycles individually.

From this initial wind field we compute the meridionally varying part of the initial temperature field following the thermal wind balance approach used by Polvani and Esler (2007). The meridionally constant part of the (potential) temperature field is specified by the profile  $\theta(z)$ , which is constructed by solving Equation A4 for given (horizontally constant) static stability  $N^2$   
 565 and surface potential temperature  $\theta_{sf_c}$ .

$$N^2(z) = (g/\theta)\partial_z\theta, \quad (\text{A4})$$

with gravitational acceleration  $g$ . The imposed profile of  $N^2(z)$  is defined by Equation A5 and consists of two regions of constant static stability ( $N_T^2$  and  $N_S^2$ , corresponding to troposphere and stratosphere) with a smooth transition at height  $z_{TP}$ .

$$N^2(z) = N_T^2 + 0.5(N_S^2 - N_T^2)(1 + \tanh((z - z_{TP})/\Delta z_{TP})) \quad (\text{A5})$$

570 To trigger wave growth due to the baroclinic instability of the system we perturb the temperature field of the initial state with a vertically and meridionally confined and zonally periodic disturbance of fixed zonal wave number  $k$ . The spatial structure  $T_{pert}$  of this temperature perturbation is defined via Equation A6. Following Polvani et al. (2004) we do not introduce an equivalent balanced wind perturbation as the small imbalance of this initial perturbation only has a negligible effect on the general evolution of the flow, compared to the rapidly growing unstable modes of the system.

$$575 \quad T_{pert} = T_{max} \cos(k\lambda) \cosh(2(\phi - \phi_{pert}))^{-2} \exp((p - p_0)/(p_0 - p_{pert})), \quad (\text{A6})$$

where  $p_0 = 1000$  hPa and  $\lambda$  is longitude. Table A1 lists the physical parameters and parameter ranges used to define the different basic states used in the present paper.

*Author contributions.* PR produced the idealised model simulations, analysed the corresponding output, produced the visualisations and  
 580 wrote the paper. TB advised PR throughout this work, contributed to the interpretation of the results and improved the paper for the final version.

*Competing interests.* The authors declare that they have no conflict of interest.

**Table A1.** Physical parameters used in the the different model experiments.

Symbol	Physical meaning	Value
$u_{Tmax}$	Tropospheric jet strength	45 m/s
$z_{Tmid}$	Tropospheric jet core height	11 km
$\alpha$	Tropospheric jet depth parameter	3
$u_{Smax}$	Stratospheric jet strength	0-75 m/s
$z_{Smid}$	Stratospheric jet core height	50 km
$\Delta z_S$	Stratospheric jet depth	22 km
$\phi_S$	Stratospheric jet core latitude	60°
$\Delta\phi_S$	Stratospheric jet width	12°
$\theta_{sfc}$	Surface potential temperature	288 K
$N_T^2$	Tropospheric static stability	1.2e-4 s <sup>-1</sup>
$N_S^2$	Stratospheric static stability	5e-4 s <sup>-1</sup>
$z_{TP}$	Reference tropopause height	12.5 km
$\Delta z_{TP}$	Reference tropopause depth	3 km
$T_{max}$	Temperature perturbation magnitude	1 K
$k$	Zonal perturbation wave number	6
$\phi_{pert}$	Perturbation latitude centre	45°
$p_{pert}$	Perturbation pressure top	700 hPa
$z_\eta$	Cut-off transition height	10 km and 25 km
$\Delta z_\eta$	Cut-off transition depth	4 km



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