The role of synoptic eddies in the tropospheric response to stratospheric variability

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[1] The tropospheric response to sudden stratospheric warmings (SSWs) is analyzed in an idealized model setup regarding the respective roles of planetary-scale and synoptic-scale waves. The control model run includes a full interactive wave spectrum, while a second run includes interactive planetary-scale waves but only the time-mean synoptic-scale wave forcing from the control run. In both runs, the tropospheric response is characterized by the negative phase of the respective tropospheric annular mode. But given their different latitudinal structure, the control run shows the expected response, i.e., an equatorward shift of the tropospheric jet, whereas the response in the absence of interactive synoptic eddies is characterized by a poleward jet shift. This opposite jet shift is associated with a different planetary wave variability that couples with the zonal flow between the stratosphere and the surface. These results indicate that the synoptic eddy feedback is necessary for the observed tropospheric response to SSWs.
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1. Introduction

[2] After a sudden stratospheric warming (SSW) event, a tropospheric response can often be observed in the extratropical troposphere [*Baldwin and Dunkerton*, 2001]. This tropospheric signal generally projects onto the intrinsic tropospheric annular mode [*Thompson and Wallace*, 2000; *Baldwin and Dunkerton*, 2001], corresponding to a latitudinal shift of the zonal mean tropospheric jet. The initial tropospheric response is likely induced by a combination of effects, including an Eliassen adjustment to the stratospheric anomaly [*Haynes et al.*, 1991]. Another mechanism points to the importance of downward wave coupling associated with planetary wave reflection [*Shaw et al.*, 2010]. The subsequent tropospheric signal often persists for several weeks up to a few months, which is considerably longer than typical tropospheric annular mode decorrelation timescales [*Baldwin et al.*, 2003; *Gerber et al.*, 2008] and which indicates a potential for seasonal prediction, given the stratospheric anomaly [*Sigmond et al.*, 2013; *Mukougawa et al.*, 2009].

[3] While it is useful to think of stratosphere-troposphere coupling separately by tropospheric synoptic-scale waves and deep planetary-scale waves that are both coupling with the zonal flow [*Plumb*, 2010], the respective roles of synoptic-scale and planetary-scale waves in the tropospheric response are not resolved [*Kunz and Greatbatch*, 2013]. *Song and Robinson* [2004] point toward the importance of planetary-scale waves in the polar lower stratosphere, but they also show an important role of the synoptic eddy feedback at the location of the eddy-driven jet, similar to [*Kushner and Polvani*, 2004, hereafter KP04]. On the other hand, *Thompson et al.* [2006] indicate that a tropospheric synoptic eddy feedback may not be necessary.

[4] This study investigates the role of synoptic eddies in the downward coupling of the stratospheric anomaly and the persistence of the tropospheric anomaly by extending the KP04 approach to separate out the synoptic eddy response. Since the total eddy forcing in KP04 was applied in a zonally symmetric model, these runs did not allow for the analysis of the transient response to stratospheric warming events with respect to different wave numbers. We here extend KP04 by applying only the synoptic-scale eddy forcing of a control run to a three-dimensional spectral core model which was truncated to only include planetary-scale wave numbers.

2. Model Setup

[5] We use the Geophysical Fluid Dynamics Laboratory dry dynamical spectral core model in T42 resolution on 40 uneven sigma levels [cf. *Chen and Zurita-Gotor*, 2008]. Following the setup and definitions by *Polvani and Kushner* [2002], we use γ = 4 K/km to define a strong midwinter polar vortex, and ϵ = –10 K to induce a tropospheric asymmetry between the winter and summer hemispheres. The model runs have no seasonal cycle in order to allow for large statistics of midwinter sudden warmings and to exclude final warmings, since these tend not to project onto the tropospheric annular modes in both reanalysis [*Black and McDaniel*, 2007] as well as model studies. Zonal wave-2 topography of height 3000 m is used following *Gerber and Polvani* [2009]. The model runs have a length of 20,000 days each, of which the last 19,600 days are used for the analysis.

[6] Two model runs are compared: A model run including all resolved wave numbers (hereafter: the *full* model run), and a model run including only the zonal mean and the planetary zonal wave numbers 1, 2, and 3 (hereafter: the *truncated* model run). All smaller-scale eddies are truncated by setting the short-wave spectral coefficient for all model variables to zero at every time step, which is quite conveniently possible due to the spectral model setup.

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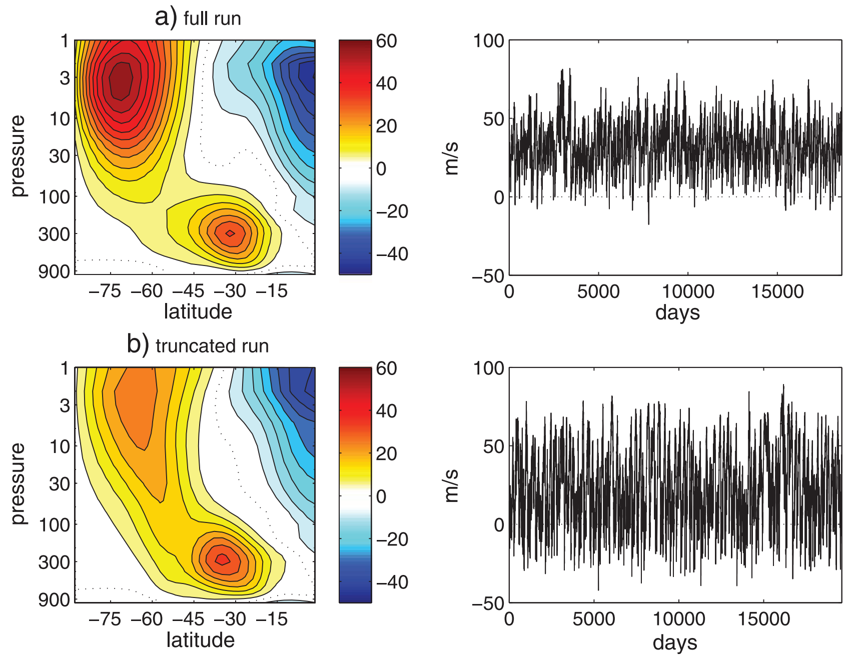

Figure 1. The zonal mean zonal wind [contour interval: 5 ms^{-1}] averaged over the entire run for (a) the full model run and (b) the truncated model run. The zero-wind line is dotted. The panels to the right show the corresponding time series at 10 hPa and 60° latitude.

[7] In order to avoid baroclinic instability of the planetary-scale waves in the truncated run [cf. *Domeisen and Plumb*, 2012] and in order to yield a troposphere that is better comparable to the full run, the time and zonally averaged synoptic (zonal wave numbers 4+) eddy forcing is diagnosed from the full run and added to the truncated run as an external forcing term in the zonal wind, temperature, and surface pressure tendency equations (more details can be found in the Appendix and in KP04).

[8] While earlier studies had found unrealistically large annular mode decorrelation timescales for this type of model [*Chan and Plumb*, 2009], the decorrelation times of this

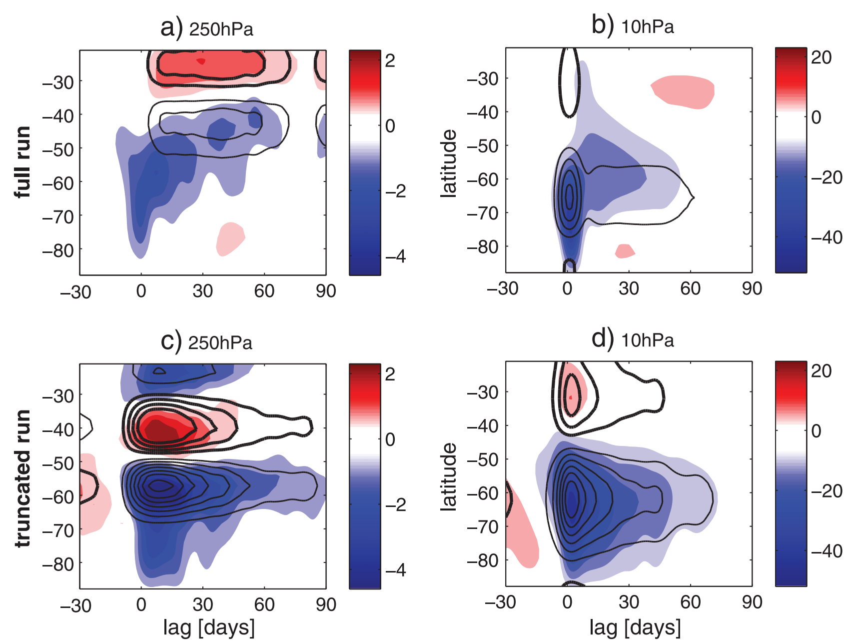

Figure 2. Composites of the 5 day running mean zonal mean zonal wind anomaly (shading) [contour interval: 0.5 ms^{-1} at 250 hPa, 5 ms^{-1} at 10 hPa] around the stratospheric sudden warming as a function of lag (with respect to the onset of the sudden warming) and latitude for (a) the full model run at 250 hPa, (b) the full model run at 10 hPa, (c) the truncated model run at 250 hPa, and (d) the truncated model run at 10 hPa. The plotted wind anomalies are significant at the 99% level (using a *t* test). The black contours (same contour interval as shading) denote the wind anomalies regressed onto the dominant empirical orthogonal function; positive patterns are printed in bold, the zero contour is omitted.

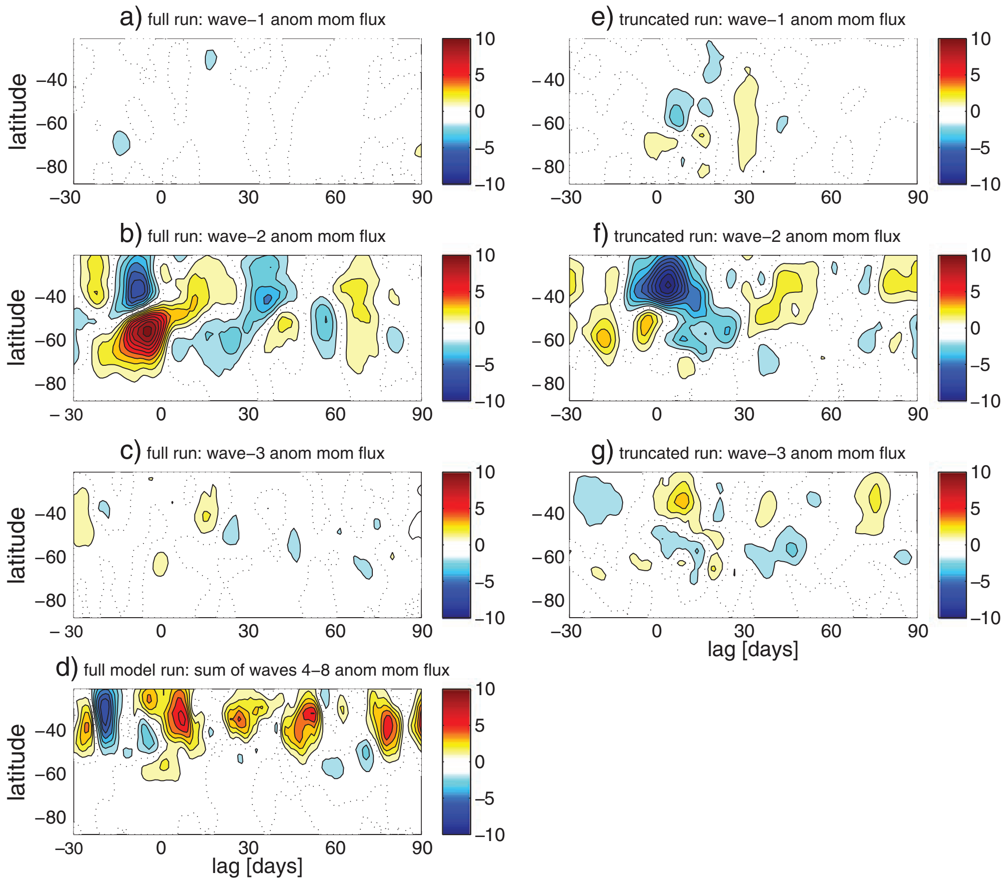

Figure 3. SSW composites of the 5 day running mean anomalous momentum flux [contour interval: 1 m^2s^{-2}] at 250 hPa as a function of lag (with respect to the onset of the sudden warming) and latitude. The dotted line denotes zero anomalies. The panels are separated by wave number for (a–d) the full and (e–g) the truncated run.

particular model setup are more realistic due to the inclusion of topography and the chosen model parameters (compare *Gerber and Polvani* [2009]). The decorrelation timescale is 40 (49) days for the full (truncated) model run at 250 hPa, and 35 (42) days at 10 hPa.

3. Results

[9] As a first step, the mean state of the troposphere is compared between the full and truncated runs. While it is not possible to exactly reproduce the troposphere of the full run in the truncated run by adding the eddy forcing, since only a time-mean eddy forcing can be added in order not to interfere with the tropospheric response to stratospheric variability, the addition of the eddy forcing nevertheless yields a comparable mean state of the troposphere: The time and zonal mean tropospheric jet is comparable in strength and location between the runs (Figure 1), and it exhibits maxima with a poleward tilt downstream of the topography for both runs (not shown).

[10] The stratospheric vortex, however, is weaker and further equatorward on average in the truncated run, while vortex variability is stronger (Figure 1). This may be due to a change in transient planetary wave propagation, which is very sensitive to the state of the tropopause [*Chen and Robinson*, 1992], i.e., despite the similar tropospheric mean state, transient wave propagation into the stratosphere may differ between the runs. While the mean state of the stratosphere differs between the truncated and the full run, we here focus on the impact on the troposphere, which is forced by the strong stratospheric anomalies that are observed in both runs. SSW events are identified (as in *Gerber and Polvani* [2009]) as the day when the principal component time series computed from zonal mean zonal wind at 10 hPa falls below two standard deviations, and events are separated by at least 45 days. This yields 93 events for the full run and 63 for the truncated run.

[11] While the evolution of the stratospheric wind deceleration is similar (Figures 2b and 2d), the tropospheric response differs considerably between the full and the truncated run (Figures 2a and 2c). In the full model run, the tropospheric jet shifts equatorward with respect to its climatological mean position (around 32°) after the onset of the SSW, consistent with observations [e.g., *Baldwin and Dunkerton*, 2001]. In the truncated run, however, the jet strengthens poleward and weakens equatorward of the climatological jet.

[12] These different responses can partly be explained from a zonal mean perspective: For both runs, the tropospheric response can be described by the dominant tropospheric mode of the respective model run (black contours in Figures 2a and 2c). The tropospheric annular mode of each run, however, exhibits a different latitudinal structure. While the stratospheric dominant mode shows a maximum at the location of the polar vortex for both runs indicating a strengthening/weakening pattern of the vortex,

the dominant tropospheric mode exhibits its node at the location of the climatological jet for the full run, while the truncated run has its node close to the latitude of the peak in the topography (45°). Deducing the tropospheric response from the stratospheric mode indicates that a weakening of the stratospheric polar vortex goes along with a negative phase of the respective tropospheric mode, corresponding to a strengthening of the tropospheric winds equatorward of the node of the tropospheric mode.

[13] From a more physical point of view, the difference between these runs indicates that planetary-scale waves interact differently with the mean flow than synoptic eddies. Interestingly, *Hitchcock et al.* [2013] also find opposing effects of synoptic and planetary-scale waves on the eddy-driven jet in response to SSWs in a different idealized model (Figures 5 and 11 of their paper). In order to examine the behavior of the different waves, Figure 3 shows anomalous momentum fluxes separated by wave number at a height of 250 hPa for the full and the truncated runs. For both runs, wave-2 is dominant among the planetary-scale waves, showing convergence at the location of the topography before the onset of the SSW, as expected from the increased wave-2 amplitude which induces the sudden warming (not shown). After the onset of the SSW, for the full run, the tropospheric jet shift is maintained by anomalous equatorward synoptic eddy momentum fluxes for over 3 months. This is consistent with observations showing that tropospheric jet variability is maintained by synoptic waves traveling meridionally away from their source region [*Lorenz and Hartmann*, 2003]. The planetary-scale momentum fluxes, on the other hand, decay in response to the equatorward-shifted jet, as was found in *Hitchcock et al.* [2013]. In the truncated run, the anomalous synoptic eddy fluxes are zero by construction, and the response is dominated by wave-2 momentum fluxes. The tropospheric response is, however, not maintained and decays with a timescale corresponding to the annular mode decorrelation timescale, yielding a different behavior of wave-2 momentum fluxes as compared to the control run. The difference between the two runs highlights the importance of synoptic wave variability in organizing the tropospheric jet, as found in observations [e.g., *Lorenz and Hartmann*, 2003].

4. Discussion

[14] In summary, sudden warmings can be observed both in the absence and presence of synoptic eddies in an idealized model simulation, i.e., in both the truncated and the full model run. The sudden warming is associated with a tropospheric convergence of wave-2 momentum flux around the topography for both runs.

[15] While for both runs, the tropospheric response can be described by the intrinsic tropospheric mode, this mode is represented by a different latitudinal structure. In the full model, the variability corresponds to a latitudinal shift about the location of the climatological jet, while in the truncated run, the signal is represented by a latitudinal shift about the topography. This yields an equatorward shift of the tropospheric jet for the full run, but a poleward shift for the truncated run as a response to the SSW, although in both cases, the tropospheric response to SSWs is characterized by the negative phase of the respective tropospheric mode.

[16] These results indicate that while planetary waves in the absence of interactive synoptic eddies are able to induce a tropospheric response to a stratospheric forcing, the equatorward shift of the tropospheric jet following a SSW, which is observed in reanalysis and idealized models, cannot be interpreted as a simple Eliassen response to the stratospheric event, but that the dynamics controlling the observed jet shift are linked to the synoptic eddy momentum fluxes. Further research will have to focus on the role of planetary-scale waves in the vertical coupling of the atmosphere [e.g., *Shaw et al.*, 2010].

Appendix A: Calculation of the Synoptic Eddy Forcing

[17] The method can be illustrated using an advection equation with a damping term

$$\frac{\partial q}{\partial t} = -u \cdot \nabla q - k(q - q_{\mathrm{eq}}) \equiv F(u, q), \tag{A1}$$

where q is a tracer, k is a damping rate, and q_{eq} is a prescribed, time-independent, and zonally symmetric equilibrium profile of the tracer. $F(u, q)$ is an operator for the instantaneous local tendency of q associated with advection and damping. Unlike the KP04 method, we derive the eddy forcing from the instantaneous fields rather than the time-averaged fields. We apply the tendency operator $F(.)$ to the zonal mean terms

$$\overline{F(\overline{u}, \overline{q})} = -\overline{u} \cdot \nabla \overline{q} - k(\overline{q} - q_{\mathrm{eq}}) \tag{A2}$$

and then to the zonal means plus the synoptic eddy component,

$$\overline{F(\overline{u} + u^s, \overline{q} + q^s)} = -\overline{u} \cdot \nabla \overline{q} - k(\overline{q} - q_{\mathrm{eq}}) - \overline{u^s \cdot \nabla q^s} . \tag{A3}$$

Here overbars denote the zonal means, and the superscript s denotes the synoptic eddy component (wave numbers 4+) of the total field. The synoptic eddy forcing can be obtained from the difference between (A2) and (A3) to yield

$$\overline{u^s \cdot \nabla q^s} = \overline{F(\overline{u}, \overline{q})} - \overline{F(\overline{u} + u^s, \overline{q} + q^s)} . \tag{A4}$$

The tendency operator $F(.)$ is obtained by integrating the primitive equation model forward by one time step using instantaneous daily zonal and meridional winds, temperature, and surface pressure. We first calculate the tendencies for the zonal mean fields and then compute the tendencies for zonal means plus the synoptic eddy component. The difference of the two yields the instantaneous synoptic eddy forcing in equation (A4). Using the primitive equation model for the tendency calculation ensures that the eddy forcings are consistent with the horizontal and vertical discretization of the numerical model as well as the topography. The time-averaged synoptic forcing is then used in the truncated run with interactive planetary-scale waves only.

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