

Maintenance of the Stratospheric Structure in an Idealized General Circulation Model

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ABSTRACT

This work explores the maintenance of the stratospheric structure in a primitive equation model that is forced by a Newtonian cooling with a prescribed radiative equilibrium temperature field. Models such as this are well suited to analyze and address questions regarding the nature of wave propagation and troposphere–stratosphere interactions. The focus lies on the lower to midstratosphere and the mean annual cycle, with its large interhemispheric variations in the radiative background state and forcing, is taken as a benchmark to be simulated with reasonable verisimilitude. A reasonably realistic basic stratospheric temperature structure is a necessary first step in understanding stratospheric dynamics.

It is first shown that using a realistic radiative background temperature field based on radiative transfer calculations substantially improves the basic structure of the model stratosphere compared to previously used setups. Then, the physical processes that are needed to maintain the seasonal cycle of temperature in the lower stratosphere are explored. It is found that an improved stratosphere and seasonally varying topographically forced stationary waves are, in themselves, insufficient to produce a seasonal cycle of sufficient amplitude in the tropics, even if the topographic forcing is large. Upwelling associated with baroclinic wave activity is an important influence on the tropical lower stratosphere and the seasonal variation of tropospheric baroclinic activity contributes significantly to the seasonal cycle of the lower tropical stratosphere. Given a reasonably realistic basic stratospheric structure and a seasonal cycle in both stationary wave activity and tropospheric baroclinic instability, it is possible to obtain a seasonal cycle in the lower stratosphere of amplitude comparable to the observations.

1. Introduction

In studies of tropospheric dynamics it has been found to be invaluable to use, in addition to comprehensive general circulation models, more idealized models that represent the thermodynamic forcing as a simple relaxation back to some prescribed radiative equilibrium temperature—a procedure commonly referred to as a “Newtonian cooling.” Such models allow the isolation of important processes without unnecessary complicating factors. One might expect that a similar procedure would work as well, or better, in the stratosphere because the dynamics and the thermodynamic forcing are in some ways simpler (water vapor has virtually no influence, for example) so that, one might hope, an idealized radiative forcing would be sufficient for many purposes. However, the stratosphere has other complications—among

other things it has a pronounced annual cycle and is strongly influenced by wave forcing from below—and it has proven quite difficult to obtain a realistic stratospheric structure using idealized forcing, especially in the tropics (Polvani and Kushner 2002; Gerber and Polvani 2009; Gerber 2012). In this paper we build on this important prior work and propose a thermodynamic forcing that overcomes some, but not all, of the problems previously encountered. This forcing then enables us to better examine the maintenance of the structure of the stratosphere and, in particular, the dynamics of the tropical stratospheric seasonal cycle.

As noted, temperatures in the lower stratosphere show a pronounced annual cycle, both in the tropics and high latitudes. To first order, the annual cycle at high latitudes in both hemispheres is a consequence of the large annual cycle in insolation. However, insolation in the tropics is more or less constant, and the observed annual cycle in tropical lower-stratospheric temperatures, broadly in phase with that at northern high latitudes, perforce requires a dynamical mechanism. Similarly, the fact that the Antarctic polar vortex is colder than the

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Arctic counterpart is predominantly a consequence of stratospheric dynamics; the reduction of ozone mixing ratios as a consequence of ozone depletion is only a secondary process that reinforces the existing asymmetry between the hemispheres. Throughout most of the year, the lowest stratospheric temperatures are observed in the tropical tropopause layer (TTL), except over Antarctica, where temperatures are lower than in the TTL. One consequence of this temperature structure is the regulation of water vapor entering the stratosphere at the tropical tropopause.

The prevailing explanation (e.g., Yulaeva et al. 1994; Rosenlof 1995) for this pattern of stratospheric variability is that the landmass distribution favors stationary planetary-scale disturbances over the NH, which can propagate into the stratosphere during hemispheric winter when the prevalent stratospheric zonal winds are westerly (Charney and Drazin 1961). The dissipation of the waves in the stratosphere forces upwelling equatorward and a downwelling motion poleward of the latitudes of the wave breaking, where the integrated momentum deposition above any given position determines the strength of the circulation. This “downward control” (Haynes et al. 1991; Holton et al. 1995; Plumb 2002) offers an elegant explanation of how midlatitude planetary-scale wave drive can induce a global-scale meridional overturning circulation in the stratosphere, and successfully explains many aspects of mid- and high-latitude stratospheric dynamics. But the nature of the upwelling in the tropics, where downward control cannot be applied owing to the smallness of the Coriolis force, remains the subject of debate (Plumb and Eluszkiewicz 1999; Scott 2002; Plumb 2002; Semeniuk and Shepherd 2001). For instance, Chen and Sun (2011) find a clear correlation between midlatitude wave forcing and the tropical annual cycle, but they concentrate on the upper stratosphere, and the resulting amplitude in lower stratospheric temperature is much weaker than observed. Randel et al. (2002) argue that extratropical wave forcing is essential to tropical variability. Conversely, tropical upwelling into the lower stratosphere may be forced by quasi-stationary tropical waves (Boehm and Lee 2003), and the annual cycle in tropical upwelling may also be a consequence of tropical waves forcing less upwelling during boreal summer when the convective heating maximum is displaced far northward (Kerr-Munslow and Norton 2006; Norton 2006). Birner and Bönisch (2011) analyzed results from a chemistry–climate model (CCM) and reanalysis data and report a distinction between a deep branch of the circulation, driven by midlatitude planetary-scale wave breaking in the middle atmosphere, and a shallow branch, more linked to subtropical synoptic and planetary wave drive in the

lower stratosphere, similarly to the schematic of Plumb (2002).

The aim of this work is to assess the sensitivities of stratospheric dynamics with an idealized dry general circulation model (GCM), which represents an efficient tool for testing basic mechanisms. We first analyze the ability of a range of dry model configurations to recover the leading-order variability of lower-stratospheric dynamics. We focus on the seasonal cycle of temperatures both because it is a well observable quantity (the residual circulation can be only indirectly inferred), and because of the importance of temperatures for stratospheric water and chemistry (e.g., Fueglistaler et al. 2009; Mote et al. 1996).

Section 2 introduces the dry dynamical core and an overview of frequently used setups, followed by a discussion of their respective stratospheric structure in section 3. We then introduce a new way of defining the stratosphere in an idealized model based on radiative transfer calculations in section 4, and show the resulting dynamic structure. Section 5 explores the role of synoptic versus orographic wave forcing in the stratosphere. We discuss our results and conclude in section 6.

2. The stratosphere in an idealized model

a. General model setup

All model integrations shown in this paper are based on the Geophysical Fluid Dynamics Laboratory’s (GFDL’s) spectral dynamical core with horizontal resolution T42 and 40 hybrid levels in the vertical, spaced as in Polvani and Kushner (2002). Of these, 24 levels are above 100 hPa. From 0.5 hPa upward, the dynamics are damped with a sponge layer (Rayleigh friction) with a time scale of 0.5 days at the top layer (Polvani and Kushner 2002). Similarly, Rayleigh surface drag extends below $\sigma = p/p_s = 0.7$, with a time scale of 1 day at the bottom layer. These are frequently used and are as described in Held and Suarez (1994). The model is forced with a Newtonian cooling term

$$Q = -\frac{T - T_e}{\tau}, \quad (1)$$

with the absolute temperature T , relaxation temperature T_e , and time scale τ . The exact forms of T_e and τ define what we call a model setup. Table 1 provides an overview of the variants of model setups discussed here. Common to all setups is that T_e and τ are zonally symmetric, and the troposphere follows closely the initial setup of Held and Suarez (1994). Explicitly,

TABLE 1. Summary of explored setups. In the first column, abbreviations are given that will be used throughout this work. The second column gives differences to the standard HS94 setup, with ε and d defined in Eq. (4). The third column gives parameters for the stratosphere, the fourth column gives the NH mountain height, and the fifth column states whether (T_e, τ) depend on time.

Setup	Troposphere	Stratosphere	Topography	Evolution
HS94	$\varepsilon = 10 \text{ K}, d = 0$	$T_e = 200 \text{ K}, \tau = 40 \text{ days}$	None	None
PK02	$\varepsilon = 10 \text{ K}, d = 0$	U.S. Standard Atmosphere + polar vortex ($\gamma = 4 \text{ K km}^{-1}$), $\tau = 40 \text{ days}$	None	Perpetual January
GP09	$\varepsilon = 10 \text{ K}, d = 0$	U.S. Standard Atmosphere + polar vortex ($\gamma = 4 \text{ K km}^{-1}$) $\tau = 40 \text{ days}$	Wave-2, 3 km, sinusoidal	Perpetual January
AN	$\varepsilon = 10 \text{ K}, d = 0$	Radiative (T_e, τ) above 100 hPa	Wave-2, 4 km, Gaussian	Perpetual January
AS	$\varepsilon = 10 \text{ K}, d = 0$	Radiative (T_e, τ) above 100 hPa	None	Perpetual January
A	AN for January, AS mirrored around equator.			
B	$\varepsilon = 10 \text{ K}$	Radiative (T_e, τ) above 100 hPa	Wave-2, 4 km, Gaussian	Full seasonal cycle
C	$\varepsilon_{\text{SH}} = 10 \text{ K}$ $\varepsilon_{\text{NH}} = 40 \text{ K}$	Radiative (T_e, τ) above 100 hPa	Wave-2, 4 km, Gaussian	Full seasonal cycle
D	$\varepsilon_{\text{SH}} = 10 \text{ K}$ $\varepsilon_{\text{NH}} = 40 \text{ K}$	Radiative (T_e, τ) above 100 hPa	None	Full seasonal cycle

$$T_e(\varphi, p, d) = \begin{cases} \max\{T_T, [T_0 - \delta T(\varphi, p, d)](p/p_0)^\kappa\} & p \geq p_t \\ T_e^{\text{strat}}(\varphi, p, d) & p < p_t, \end{cases} \quad (2)$$

where φ is latitude, p is pressure, d is the day of the year, and

$$\delta T(\varphi, p, d) = \delta_y \sin^2 \varphi + \delta_z \log(p/p_0) \cos^2 \varphi + \tilde{\varepsilon}(\varphi, d) \sin \varphi \equiv T_e^{\text{sym}}(\varphi) + \tilde{\varepsilon}(\varphi, d) \sin \varphi, \quad (3)$$

and $T_0 = 315 \text{ K}$, $p_0 = 1000 \text{ hPa}$, $\kappa = 2/7$, $\delta_y = 60 \text{ K}$, $\delta_z = 10 \text{ K}$, and $p_t = 100 \text{ hPa}$. We set T_T to either 200 or 216 K to match the lower boundary of T_e^{strat} . Also, T_e^{sym} is north–south symmetric, and we introduce a time- and hemisphere-dependent asymmetry in the troposphere with

$$\tilde{\varepsilon}(\varphi, d) \equiv \varepsilon(\varphi) \cos(2\pi d/365) = \cos(2\pi d/365) \times \begin{cases} \varepsilon_{\text{NH}}, & \varphi > 0 \\ \varepsilon_{\text{SH}}, & \varphi < 0 \end{cases}. \quad (4)$$

For perpetual winter runs, we set $d = 0$. This north–south gradient in T_e (with $d = 0$ and $\varepsilon_{\text{NH}} = \varepsilon_{\text{SH}}$) has been used by all earlier model setups discussed in this work (Polvani and Kushner 2002; Gerber and Polvani 2009; Chan and Plumb 2009). The so-computed (T_e, τ) pair can then conveniently be written into a climatological input file for the dry dynamical core. In this paper, we will mainly be concerned with the form of $T_e^{\text{strat}}(\varphi, p, d)$, but the north–south asymmetry in Eq. (4) will also be important for what follows.

For discussing our results, it will be convenient to introduce a notation for the dynamical contribution to the temperature field $\mathcal{T}$, defined by

$$\mathcal{T} \equiv T - T_e. \quad (5)$$

Note that with this notation, the quasi-geostrophic, steady-state diabatic zonal mean residual vertical velocity $\bar{w}^*$ is

$$\bar{w}^* = -\frac{R}{N^2 H} \frac{\mathcal{T}}{\tau}, \quad (6)$$

where R is the gas constant of air, N is the buoyancy frequency, and H is the scale height. Over large parts of this work, we will be concerned with the seasonal cycle in the tropical lower stratosphere, where T_e is nearly constant in time (shown below), and the easily measured and available temperature T is directly related to upwelling through Eq. (6).

All results from model integrations with time-constant equilibrium temperature structure are averages over 2000 days, after a model spinup time of 500 days. These long integrations ensure statistically robust results. The seasonality is evaluated from the differences in the steady-state model results for setups with and without orographic forcing. The seasonally variable integrations of Table 1 are performed with a seasonally varying Newtonian cooling term. For these calculations, the climatological mean annual cycle is determined from the monthly means averaged over the last 6 years after a spinup of a few years to achieve statistical steady state. We have checked convergence with up to 10 years of integration time, and did not find any significant differences.

Figure 1 shows some climatological results from the Interim European Centre for Medium-Range Weather

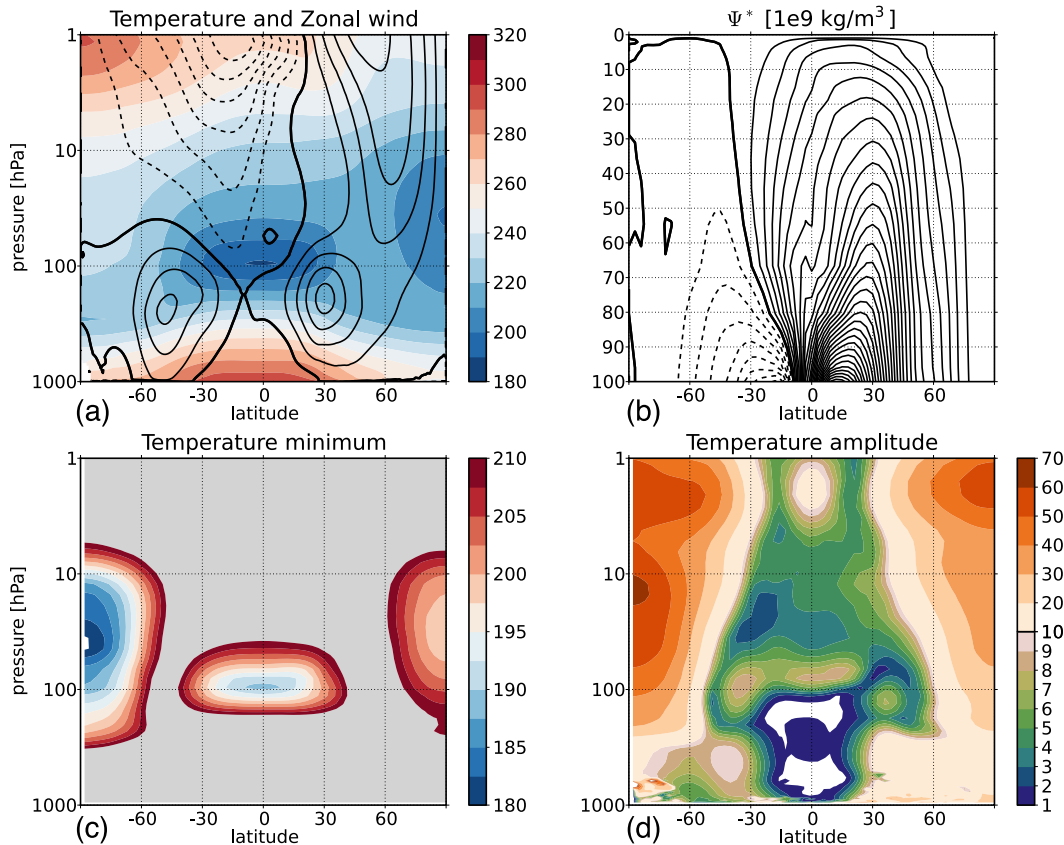


FIG. 1. ERA-Interim climatology. (a) Zonal mean January temperature and zonal wind (contour interval 10 ms^{-1}), (b) residual streamfunction (contour interval $0.5 \times 10^9 \text{ kg m}^{-3}$), (c) minimum monthly temperature, and (d) maximum monthly temperature amplitude in the annual cycle. Solid lines denote positive, dashed lines denote negative, and thick solid lines denote zero values. Values in the color bars are in kelvins.

Forecasts (ECMWF) Re-Analysis (ERA-Interim) (Dee et al. 2011), which will be used throughout this article for comparing model results to observations. Figures 1a and 1b show climatological January zonal mean temperatures (colors) and zonal wind (contours), as well as residual streamfunction. Figures 1c and 1d are based on monthly climatology and display the minimum monthly zonal mean temperature and the difference between the maximum and the minimum temperatures in the annual cycle.

b. The Held–Suarez (HS94) setup

A basic state for Newtonian cooling in the troposphere has been proposed by Held and Suarez (1994, hereafter the HS94 setup). This model relaxes the temperature to an analytical latitude–pressure structure, with a relaxation time of 40 days almost everywhere, the only exception being a shortening to 4 days in the tropics close to the surface. The stratosphere is given a constant equilibrium temperature of $T_T = T_e^{\text{strat}}(\varphi, p, d) = 200 \text{ K}$.

Figures 2a and 2b show the results for the HS94 setup, showing the same fields as Figs. 1a and 1b. The model produces a hemispheric-scale residual circulation well into the lower stratosphere despite the absence of a polar night jet—a feature we will further discuss later. The direct consequence of the diabatic residual circulation is that temperatures in the upwelling branch in the tropics are below, and in the downwelling branch above the equilibrium temperature (Fig. 2b). In the HS94 setup, temperatures at the tropical tropopause reach a minimum of about 190 K, giving $T = -10 \text{ K}$. Figures 2a and 2b show that the north–south asymmetry in the tropospheric equilibrium temperature field [Eq. (4)] induces a small hemispheric asymmetry in extratropical lower stratospheric temperatures (Fig. 2a). Note that the Northern Hemisphere–Southern Hemisphere temperature difference is less than that of the prescribed equilibrium temperature T_e ; the stronger residual circulation in the Northern Hemisphere partially counteracts the hemispheric asymmetry in T_e , as seen in the larger temperature difference $T = T - T_e$ in the Northern Hemisphere (Fig. 2b).

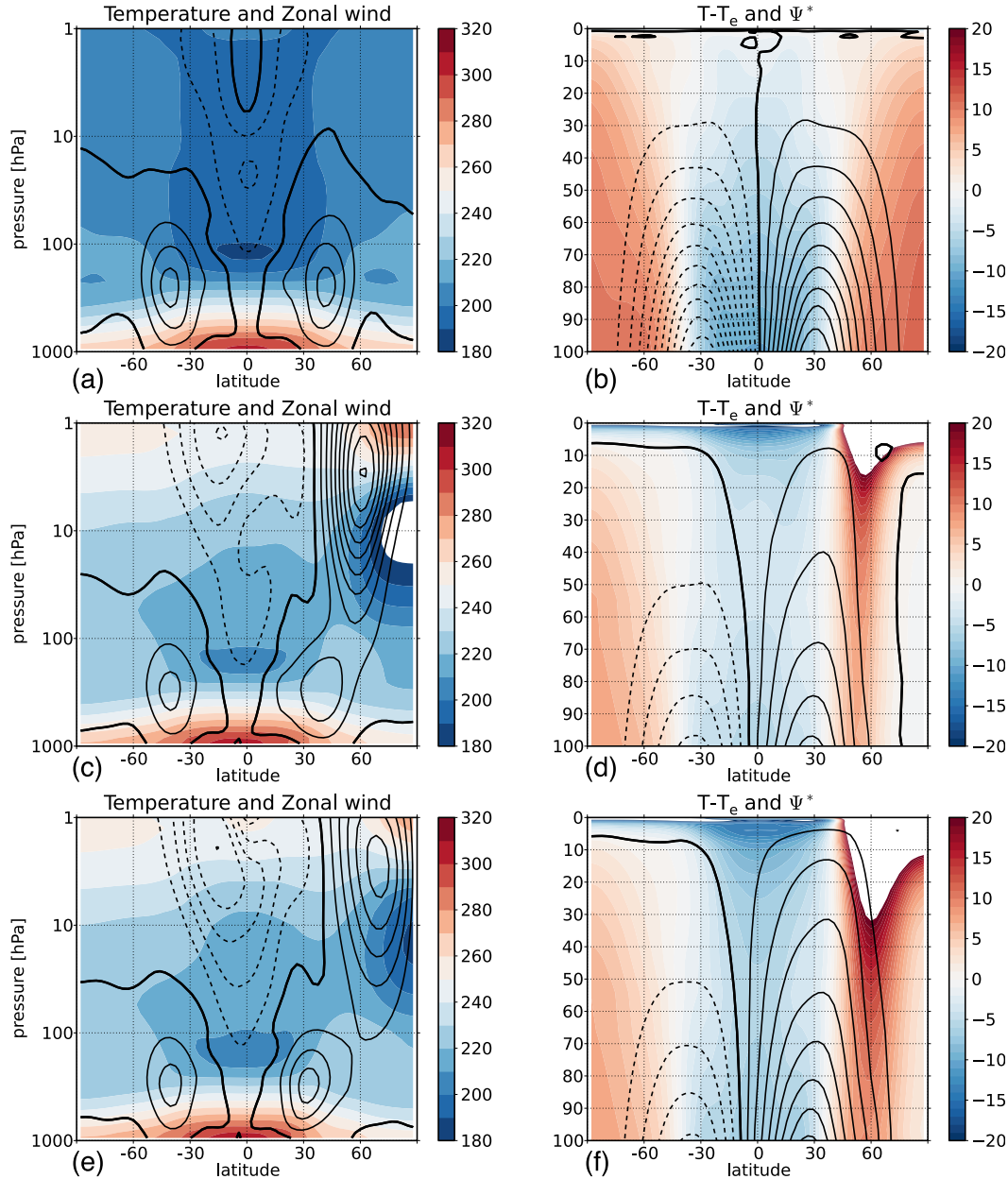


FIG. 2. Direct comparison of typical model outputs from (a),(b) HS94, (c),(d) PK02, and (e),(f) GP09. (left) Dynamic temperature (shaded colors) and zonal mean zonal wind (contours). (right) $T = T - T_e$ (colors) and the residual streamfunction (contours). Note the different range and scaling of the y axis. Contour spacing and color bars are as in Fig. 1.

c. The Polvani–Kushner (PK02) setup

The lack of a polar night jet renders the original HS94 setup ill suited for studies of the climatologically important interactions between the stratosphere and troposphere in the presence of a polar vortex. Polvani and Kushner (2002) therefore proposed a setup (the PK02 setup) based on HS94, but with a polar night region of variable strength. This setup uses the HS94 troposphere

with $T_T = 216$ K and takes above 100 hPa the analytic form $T_e^{\text{strat}}(p, \varphi) = [1 - W(\varphi)]T_{\text{US}}(p) + W(\varphi)T_{\text{PV}}(p)$, where T_{US} is the U.S. Standard Atmosphere and $T_{\text{PV}}(p) = T_{\text{US}}(100 \text{ hPa})(p/100 \text{ hPa})^{-R\gamma/g}$. The strength of the polar vortex is controlled by the lapse rate γ (K km^{-1}). The latitudinal weighting function is defined as $W(\varphi) = 0.5\{1 - \tanh[(\varphi + 50)/10]\}$.

Note that in addition to the inclusion of a polar night region, the relaxation against the U.S. Standard

Atmosphere induces a profound change in the temperature and wind structure compared to the HS94 setup. Figure 2c shows the resulting temperatures and winds for this setup with $\gamma = 4$. This model was designed primarily for studying high-latitude stratosphere–troposphere coupling, and the figure shows that indeed the mid- and high-latitude stratospheric temperature structure with this setup is more realistic (cf. Fig. 1a). But the tropical tropopause region is now substantially too warm because of the relaxation against the typical observed midlatitude temperature profile (which the U.S. Standard Atmosphere represents). Indeed, the tropopause is substantially lower in altitude than observed, and the subtropical jets do not extend high enough.

The factor γ allows variations of the strength of the polar vortex, and with $\gamma = 4$ this simulation has a relatively strong, isolated vortex. With this configuration, the vortex region is close to radiative equilibrium, and the downwelling branch of the residual circulation (Fig. 2d) is at the edge of the polar vortex.

d. The Gerber–Polvani (GP09) setup

Gerber and Polvani (2009, hereafter the GP09 setup) use the same Newtonian cooling setup as PK02 but include surface topography to study the effects of planetary scale, quasi-stationary waves on the stratospheric circulation. A plausible setup emulating the Northern Hemisphere is a wave-2 topography, with the height maxima centered at 45° latitude, and the height of the “mountains” h as an adjustable parameter to modify the wave activity at these wavenumbers.

Figures 2e and 2f show the results for this setup with $\gamma = 4$ and $h = 3000$ m, considered the most realistic combination of parameters for high-latitude stratospheric variability (GP09). The figures show that with this setup, the stratospheric residual circulation is stronger than in the PK02 setup in the hemisphere with the mountain, as expected from the larger wave activity at wavenumbers that can penetrate into the stratosphere. Also, the polar vortex is now weaker (the zonal winds decrease by about $40\text{--}45\text{ m s}^{-1}$) as a consequence of the stronger momentum deposition. Consistent with the stronger circulation, the dynamical contribution to the temperature T is more important in the GP09 than in the PK02 setup. Thus, the polar vortex is warmer than in the PK02 simulation, closer to the observed temperature of Fig. 1a. In the tropics, temperatures are cooler than without orographic forcing but are still considerably warmer than in reanalysis. The planetary-scale waves of the GP09 setup have only little bearing on upwelling in the lowest tropical stratosphere (see Fig. 4, discussed in the next section).

3. Discussion of the stratospheric structure of the HS94, PK02, and GP09 setups

All three model setups evaluated do not have a seasonally evolving prescription of Newtonian cooling in the stratosphere. To investigate the implied seasonality of these model setups, we will compare perpetual winter simulations, once with and once without topography in the Northern Hemisphere. In the context of lower stratospheric dynamics, two aspects of the model results shown in Fig. 2 stand out. First, the strongest residual circulation (in the lower stratosphere) is obtained with the HS94 setup that does not have a polar vortex. Second, whereas there are clear differences in the polar vortex strength, these integrations do not show a strong difference in the tropics between the setups, not even when comparing integrations with flat topography (Figs. 2c,d) with integrations including wave-two surface topography in the winter hemisphere (Figs. 2e,f). In the following, we first focus on differences in wave activity and upward propagation of waves that largely explain these features. In a second step, we evaluate the model results with respect to their ability to capture the key aspects of the stratospheric temperature field as shown in Fig. 1.

a. Wave activity

Figure 3a depicts the latitudinal structure of the total eddy heat flux at 500 hPa (dashed lines), indicative of tropospheric wave activity, and in the lower stratosphere at 70 hPa (solid lines). The temperature perturbations are further decomposed into wavenumbers 1–4 in Fig. 3b and wavenumbers 5–64 in Fig. 3c. The red lines labeled “A” will be described later.

A first observation is the (expected) symmetry break between the two hemispheres in the PK02 and GP09 setups, whereas the HS94 case is almost perfectly symmetric. The latter has the largest maxima of the three cases at both pressure levels. Among the three setups, PK02 and GP09 include a weaker meridional temperature gradient (and therefore weaker synoptic wave activity), as the minimum T_e [parameter T_T in Eq. (2)] has a higher value of 216 K compared to the 200 K in HS94. This is due to the stratosphere being relaxed to the observed U.S. Standard Atmosphere in PK02 and GP09, which dictates that temperature at 100 hPa. As mentioned earlier, the subtropical jets in the HS94 setup reach higher altitudes than in the other setups (see Fig. 2), allowing synoptic-scale waves to propagate farther upward than in the other setups. As expected, the PK02 and GP09 setups are identical in the (summer) SH, but the additional orographic forcing in the (winter) NH influences the position of the eddy-driven jet, and adds

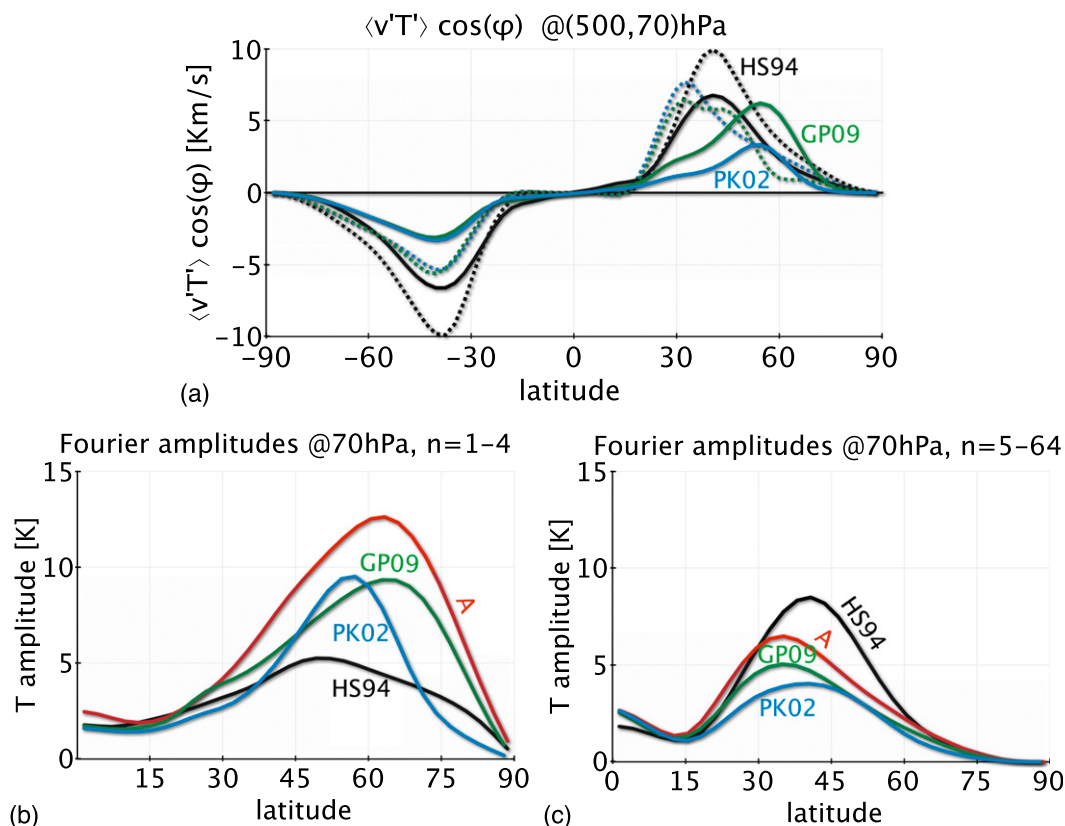


FIG. 3. Meridional heat fluxes for the model setups from Fig. 2. (a) $\overline{v'T'} \cos \varphi$ at 500 (dashed) and 70 hPa (solid), and Fourier amplitudes of temperature perturbations for (b) zonal wavenumbers 1–4 and (c) wavenumbers 5–64. Labels denote integrations listed in Table 1.

more planetary-scale waves at 70 hPa. Clearly visible is the latitudinal deviation of the heat fluxes poleward with height (from dashed to solid lines), as the polar vortex creates a waveguide around 60° , which represents an important improvement compared to the simple upward propagation of the HS94 integration.

b. Temperature

Figure 4 shows the temperature minimum and amplitude for the GP09 setup (the most realistic of the three setups), which may be directly compared to the observations shown in Fig. 1. Here, we compared the GP09 perpetual January simulations (Fig. 2e,f) to the PK02 perpetual January simulations (Fig. 2c,d), such that any nonzero value in Fig. 4 can be attributed to the presence or absence of topography. To construct the plots similar to the seasonally varying scenarios, we have mirrored the PK02 simulation around the equator. The temperature minimum and amplitude shown in Fig. 4 compare two perpetual simulations and represent orographically driven changes in circulation in the GP09 model.

With this setup, the temperatures over the Arctic are rather low, and even though they are somewhat higher

than the Antarctic temperatures, they are much colder than the tropics. These temperature minima are also quite high in altitude—around 10 hPa. The tropical cold point is very warm and too low—around 150 hPa (as opposed to about 90 hPa in observations). Similar to Chen and Sun (2011), we find that the stronger circulation due to increased topography in one hemisphere leads to an annual cycle in the mid- to upper stratosphere, but Fig. 4 makes clear that this seasonal cycle is much weaker. In addition, the seasonal cycle in the lower tropical stratosphere, of about 8 K around 80 hPa in the reanalysis, is almost completely absent (~ 1 K) in this setup. This result is consistent with the findings in Gerber (2012).

In summary, the GP09 setup qualitatively recovers the basic aspects such as a warmer polar vortex in the Northern Hemisphere and an annual cycle in tropical lower-stratospheric temperatures. But it is quantitatively far from the observed climatology, especially in the tropics, where seasonal temperature variations are dominated by dynamical T , not radiative T_e (Yulaeva et al. 1994; Fueglistaler et al. 2011). The question is whether this quantitative mismatch is due to an unrealistic

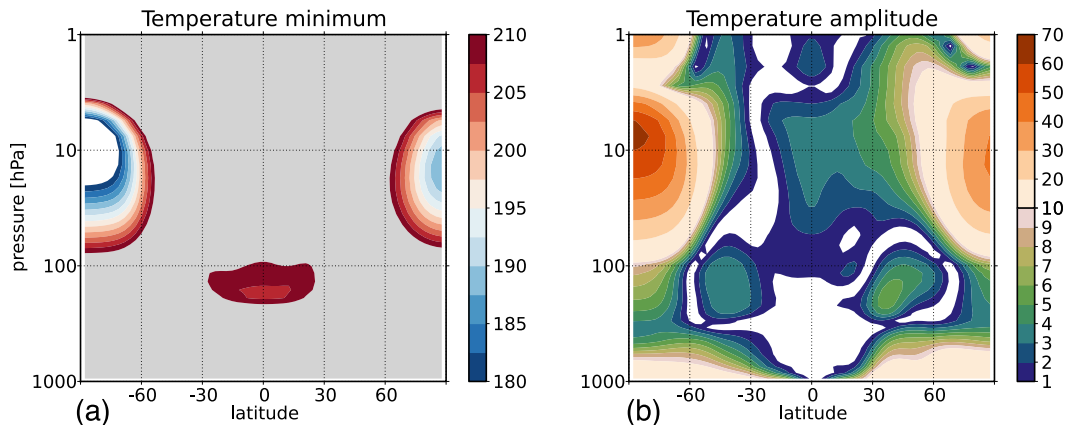


FIG. 4. As in Figs. 1c and 1d, respectively, but with the GP09 stratosphere and topography (integration GP09).

Newtonian cooling setup in the stratosphere, or whether this indicates that the model misses essential processes. In the following, we propose an improved stratospheric Newtonian cooling setup, based on radiative transfer calculations.

4. Newtonian cooling in the stratosphere based on radiative transfer calculations

We have shown that the three setups discussed so far cannot recover key aspects of the lower-stratospheric structure. In line with a large body of previously published literature, we observe that differences in the zonal wind field in the upper troposphere–lower stratosphere region have a profound impact on the resulting stratospheric circulation. This wind field, in turn, is largely determined by the prescribed equilibrium temperature field, and the question arises to what extent a more realistic Newtonian cooling setup would improve the model's stratosphere. For instance, in the above discussed setups, the relaxation time is set to 40 days everywhere. In the troposphere, this value (together with the corresponding T_e profile) has proven to be a good choice for a realistic climatology, and in particular eddy heat fluxes (HS94). In the stratosphere, however, there is no physical reason to assume the same time scale would be appropriate. Rather, the importance of latitudinally and temporally dependent τ is well documented (e.g., Kiehl and Solomon 1986; Newman and Rosenfield 1997). Not only should τ depend on latitude and time, but also altitude (Hitchcock et al. 2010; Hartmann 1981; Fels 1982; Ghazi et al. 1985).

a. T_e and τ determined from radiative transfer calculations

The lack of substantial latent heating in the stratosphere renders a Newtonian cooling approximation to radiative heating a viable alternative to full radiative

transfer calculations. Following Hartmann (1981), we determine (T_e, τ) pairs from radiative transfer calculations, whereby first the radiative heating (divergence of radiative flux) is evaluated for the base-state profile. The base state is defined by the monthly mean zonal mean climatology from reanalysis (ERA-Interim), including temperature and water vapor. Ozone mixing ratios correspond to the Fortuin and Langematz (1994) climatology, and all other radiatively relevant trace gases are set to present-day mean values. Solar insolation is set to the diurnal mean of each month at each latitude. The radiative heating corresponding to a given state is computed offline with radiative transfer calculations, for which we use the Rapid Radiative Transfer Model (RRTM) (Mlawer et al. 1997). We denote in what follows the base state with temperature T and heating rate Q .

The second step is to perturb the (dynamic) temperature and denote the new value T' . Radiative damping depends on the vertical scale and shape of the temperature perturbation (e.g., Fels 1982), and a given set of T_e and τ is strictly valid only for a specific temperature perturbation profile applied to a specific temperature and tracer profile. In the following, we take the pragmatic approach that we consider the Newtonian cooling approximation strictly as a linearization to temperature perturbations with a specific vertical scale. We expect to obtain a reasonable result for a perturbation length of 10 km, similar to the scale of the observed annual cycle of the lower stratosphere (Fueglistaler et al. 2009). The seasonality in stratospheric dynamics and transport also implies seasonal variations in radiatively active trace gases, which cannot be recovered with a linearization to temperature perturbation. We therefore evaluate the temperature linearization for each month separately with the corresponding tracer fields.

As a consequence of the temperature perturbation, the radiative heating rate will change everywhere in the

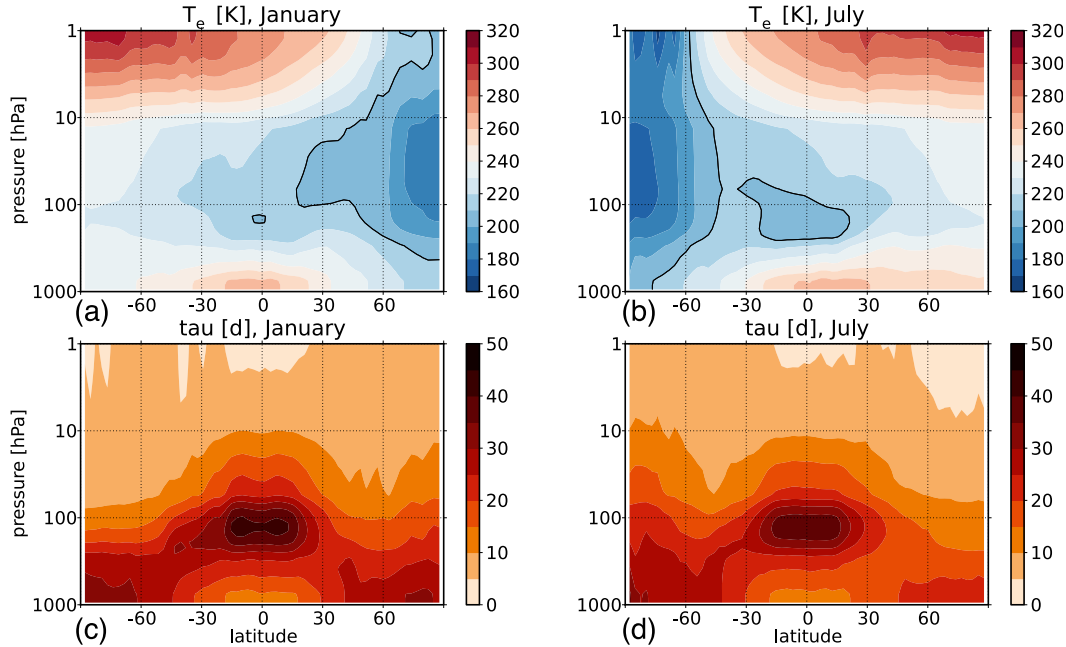


FIG. 5. (a),(b) T_e (K) and (c),(d) τ (days) for the Newtonian relaxation, corresponding to (left) January and (right) July, determined by applying a 0.1-K temperature perturbation with a vertical scale of 10 km, according to Eq. (7). For orientation, the 210-K line has been added to (a),(b).

profile. If we consider the situation at the center of the perturbation at pressure p_0 , we then have two pairs of temperature and radiative heating: $T(p_0)$ and $Q(p_0)$ for the base state, and $T'(p_0)$ and $Q'(p_0)$ for the perturbed case. For small perturbations $T - T'$ (0.1 K in our case), the response in radiative heating is approximately linear and we determine the Newtonian cooling parameters T_e and τ as

$$\tau(p_0) = -\frac{T'(p_0) - T(p_0)}{Q'(p_0) - Q(p_0)} \quad \text{and} \quad T_e(p_0) = T(p_0) + \tau Q(p_0). \quad (7)$$

We can evaluate full vertical profiles of T_e and τ by shifting the center of the perturbation p_0 (in 2.5-km steps from the bottom to 65 km in this work) and reevaluating Eq. (7) for each p_0 . Similarly, we determine $T_e(p)$ and $\tau(p)$ for each latitude and month. As a result, we can obtain spatially and temporally dependent $T_e(\varphi, p, t)$ and $\tau(\varphi, p, t)$, where φ is latitude.

The radiatively determined structure of T_e and τ cannot account for the heating in the troposphere arising from latent heat release. We therefore retain the setup of HS94 for the troposphere, and pass from one to the other at 100 hPa. We favor lower-stratospheric T_e and τ to transition smoothly into the tropospheric values of 200 K and 40 days, respectively, which represents an

important criterion for the choice of the vertical temperature perturbation scale. After a scan in vertical perturbation scales (not shown), we found that with a value of 10 km, around 100 hPa τ is a little shorter and T_e a little warmer than their HS94 counterparts. With a longer vertical perturbation, T_e becomes too cold; with a shorter perturbation, τ is too short. It is remarkable how the vertical length scale of the seasonal cycle signal in the lower stratosphere coincides well with the vertical perturbation scale required for matching the stratosphere with the HS94 troposphere. The resulting temperature and relaxation time structures are shown in Fig. 5. We note the prominent maximum of τ around the tropopause typical for this type of perturbation [a similar structure is reported by Hartmann (1981), for example]. In the upper stratosphere, τ decreases to about 4–5 days, again in accordance with previous findings.

We find that T_e compares well with Newman and Rosenfield (1997), and the polar vortex and cool tropics are clearly distinguishable. However, T_e does not steadily decrease with height over the winter pole, as one would expect for purely radiative calculations. But we are not attempting to find the radiative equilibrium temperature in the absence of circulation: Our goal is to determine a relaxation temperature and time scale representing a linear approximation to radiation around the full (i.e., measured) mean state. We argue that one can improve the climatology and sensitivity of a model

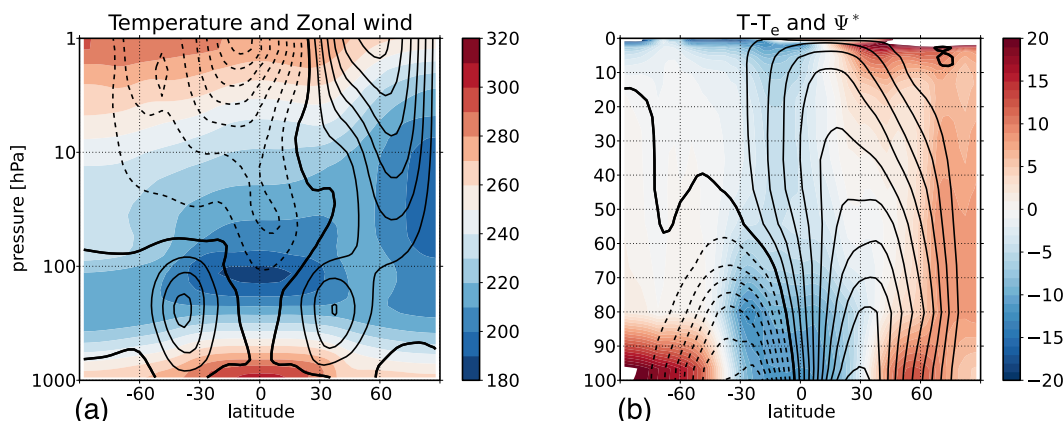


FIG. 6. As in Fig. 2, but for our new perpetual January setup (integration A). Notice how the zonal wind, temperature, and circulation are much closer to reanalysis (Fig. 1) than the earlier setups (Fig. 2).

driven by Newtonian cooling not by finding a better representation of a purely radiative, stationary state, but by damping perturbations around the mean state more realistically. As a result, our base state includes effects of resolved and unresolved wave drag in the mesosphere, yielding a warmer T_e through downwelling. This warmer T_e compared to pure radiative equilibrium assures adequate damping of perturbations, which would not be possible in such an idealized model otherwise.

b. Results based on the radiatively determined setup

We combine the above-described Newtonian cooling setup with surface topography similar to that employed in GP09 to evaluate whether the planetary-scale waves in this setup produce a more realistic stratospheric structure than in the setups analyzed in section 3. Instead of the wave-2 cosine function, we model the positive parts of the cosine in longitude with two Gaussian mountains. These mountains have the same shape as the wave-2 cosine function, less the negative topography values. This will have a somewhat smaller effect on circulation (at same mountain height), but we will compensate for that to some extent by using 4-km height instead of 3 km. We decided to use Gaussian mountains, because this possibility is already implemented in GFDL's spectral dry dynamical core, and no additional input file is necessary.

As in the previous setups, we perform perpetual January integrations—once with orographic forcing in the winter hemisphere (setup AN) and once with a flat surface (setup AS). Figures 6, 7a, and 7b are the equivalent of Figs. 2 and 4 for simulation A of Table 1. With our new setup, several key aspects of the lower stratosphere are markedly improved. The tropospheric jets reach again higher, and the winter hemisphere jet is connected to the polar vortex. This leads to increased

Eliassen–Palm (EP) fluxes, as shown by the red lines (marked “A”) in Figs. 3b and 3c. Both more synoptic- and planetary-scale waves reach the 70-hPa pressure surface compared to the HS94, PK02, and GP09 setups. As a result, the streamfunction in Fig. 6b is much stronger in the winter hemisphere compared to the other models. Also, the new setup places the tropical tropopause height and temperature closer to reanalysis. Similarly, the temperatures of the polar vortices now show a more realistic vertical structure. However, two important deficiencies remain. First, the temperature difference between the polar vortices (entirely due to differences in T in these setups) is substantially smaller than in the observations. Second, the temperature difference between the AN and AS calculations around the tropical tropopause (about 1 K, again due to differences in T) remains substantially smaller than observed (about 8 K).

The variations in the circulation have an impact on the distribution of stratospheric trace gases such as ozone and water vapor, which are important for radiative equilibrium. In addition, circulation changes also modify the base state around which we linearized the radiative effects to find T_e and τ . As a consequence, a variation in the strength of the circulation leads to a variation in radiative properties, which in our case means a difference in (T_e, τ) . This modification of T_e affects temperature and zonal winds, which in turn modify the refractive index of the stratosphere (and even the troposphere by influencing the tropospheric jets). EP fluxes are directed differently in the SH than the NH, again inducing a difference in hemispheric circulations and eventually temperature response in T . These effects cannot be included when we compare setups where the only difference is the presence or absence of orographic forcing.

To account for the seasonality in T_e , we repeat the radiative calculations described above for every month

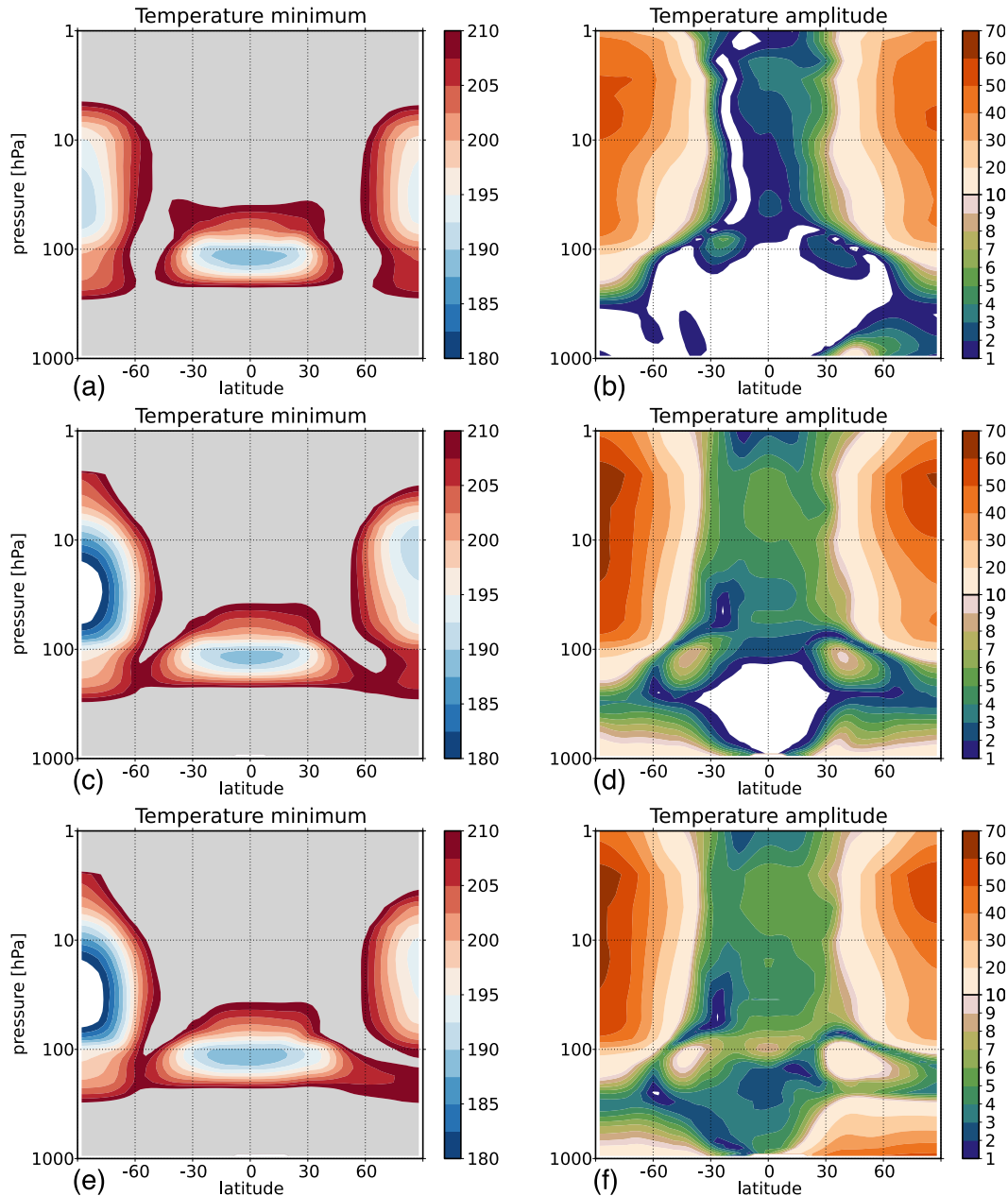


FIG. 7. Temperature minimum and amplitude, as in Figs. 1 and 4, but using our new (T_e, τ) setup. (a),(b) The seasonal cycle for perpetual January with topography (setup AN) compared to perpetual January without topography (setup AS); e.g., simulation A. (c),(d) Simulation B, including a full seasonal cycle, with a symmetric troposphere. (e),(f) Simulation C, closest to reanalysis (Fig. 1), where in addition to a full seasonal cycle, the troposphere is north-south asymmetric (see Fig. 8).

of the year, and in such a way obtain a time-dependent climatological (T_e, τ) pair (simulation B). Not only can we include the climatological differences in tracer distribution in both latitude and time, but we can also include the breakup and forming of the polar vortex in spring and autumn, which is not the case in perpetual winter simulations. Indeed, as shown in the result (Figs. 7c,d), the

polar temperatures are now closer to the reanalysis. Note that the January polar temperatures are warmer than in the perpetual January simulation, even though T_e and τ are exactly the same. This suggests that the forming and breaking up of the polar vortex in autumn and spring on one hand, and the limited duration of winter in the time-dependent case on the other hand, can have important

effects on T and thus January and July mean temperatures. The southern polar vortex is considerably cooler than its northern counterpart, just as seen in observations, and as expected from differences in tracer distributions. In addition, the minimum temperatures over the poles and the tropics have now the correct ordering. In the tropics, the annual cycle shows now a strong signal in the upper stratosphere and is about 1 K stronger than before in the TTL. This corresponds to the TTL amplitude in T_e due to the annual cycle in ozone (not shown), and is in accordance with the ozone contribution found by Fueglistaler et al. (2011).

Our results so far indicate that the variability of the extratropical stratosphere, and the tropical stratosphere above 70 hPa, is determined by the deep meridional overturning circulation, forced by midlatitude planetary wave breaking, including the associated tracer transport. This circulation induces diabatic cooling in the low latitudes and diabatic heating in the high latitudes on one hand, and enhanced tracer transport with stronger circulation on the other hand, both resulting in a warmer NH polar vortex compared to the SH polar vortex, and a seasonal cycle in tropical temperatures, mainly above 70 hPa. The TTL between 100 and 70 hPa, however, seems to be much less sensitive to these forcings.

5. The role of synoptic wave activity

Several studies suggest that tropical upwelling (and thus TTL temperature) is largely influenced by the shallow branch of the Brewer–Dobson circulation, which in turn is determined by extratropical synoptic wave activity in the troposphere (Plumb 2007; Chen and Sun 2011; Haqq-Misra et al. 2011). Indeed, as shown earlier, already in the HS94 model tropical upwelling into the stratosphere is present, and much earlier, Manabe and Mahlman (1976) obtained a reasonable seasonal cycle in the tropical tropopause temperature, even though that model missed many stratospheric processes and only had 11 levels—3 above 100 hPa. Kerr-Munslow and Norton (2006) found in their ECMWF analysis that the 100–70-hPa tropical seasonal cycle can be linked to the vertical component of the EP flux divergence and is tied to upwelling due to mass outflow just above that layer, suggesting that midlatitude wave drive is not the reason for the TTL seasonal variations.

a. North–south asymmetric troposphere

The importance of the troposphere for the lower tropical stratosphere suggests that the seasonal variability of the TTL might to a large degree be influenced by the state of the troposphere, and in particular its baroclinic activity. To test this, we let the antisymmetric part of

T_e , ε in Eq. (4), be a function of hemisphere. With this, we shape the meridional temperature gradient in a different way in the NH as compared to the SH. If we keep $\varepsilon_{\text{SH}} = 10$ K in the SH and set $\varepsilon_{\text{NH}} = 40$ K in the NH, we increase the meridional temperature gradient in the troposphere in the Northern Hemisphere winter, and relax it in summer, compared to the runs shown above. Consequently, we induce a much stronger seasonal variation in the NH than in the SH troposphere. The tropospheric relaxation temperature profiles for the two cases (symmetric and asymmetric ε) for January and July are shown in Fig. 8. This is the tropospheric setup for simulations C and D (the stratosphere being the same as in simulation B), and we will show in this section that simulation C mimics most closely the seasonal cycle found in the reanalysis climatology.

The minima and amplitudes of the monthly mean dynamic temperature climatology for simulation C is shown in Figs. 7e and 7f. Enhancing the seasonal cycle of $\partial_y T_e$ in the NH troposphere results in a much stronger seasonal cycle in TTL temperature (almost 10 K), whereas the extratropical stratosphere remains almost unchanged (cf. the middle panels of Fig. 7). These plots are remarkably similar to the corresponding ERA-Interim results of Figs. 1c and 1d. It is important to emphasize that the seasonal cycle in TTL temperature is dominated by dynamical driving (T) and is not coming from a seasonal cycle included in the setup through T_e and τ . This can be seen in Fig. 9, where we show the seasonal cycles of T_e and T for the TTL (average from 20°S to 20°N at 96 hPa) in Fig. 9a, and for the poles (averages from 70°N/S to the poles at 50 hPa) in Fig. 9b. Clearly, the tropical seasonal cycle in T is almost entirely due to dynamical effects, with T_e varying by less than 1 K owing to the seasonal cycle in ozone concentration (cf. Fueglistaler et al. 2011), and the dynamical contribution varying by 7 K in the tropical average. Over the poles (Fig. 9b), the leading-order process for the seasonal cycle in T is the change in T_e , mainly because of insolation. The difference in amplitude of the seasonal cycles in T_e between the Arctic and Antarctic (due to ozone) is about $\Delta T_e = 15$ K at 50 hPa, whereas the difference in dynamical forcing amounts to about $\Delta T = 5$ K. Thus, the differences in the strength of the seasonal cycle can be attributed to 25% dynamic and 75% ΔT_e forcing over the North and South Poles at 50 hPa, whereas it is about 85% dynamic and 15% ΔT_e in the TTL. As T_e is not purely radiative but a linearization around an observed state, the above percentages for dynamic effects probably represent lower boundaries when compared to purely radiative effects.

In the remainder of this section, we will further analyze how the state of the troposphere induces the difference

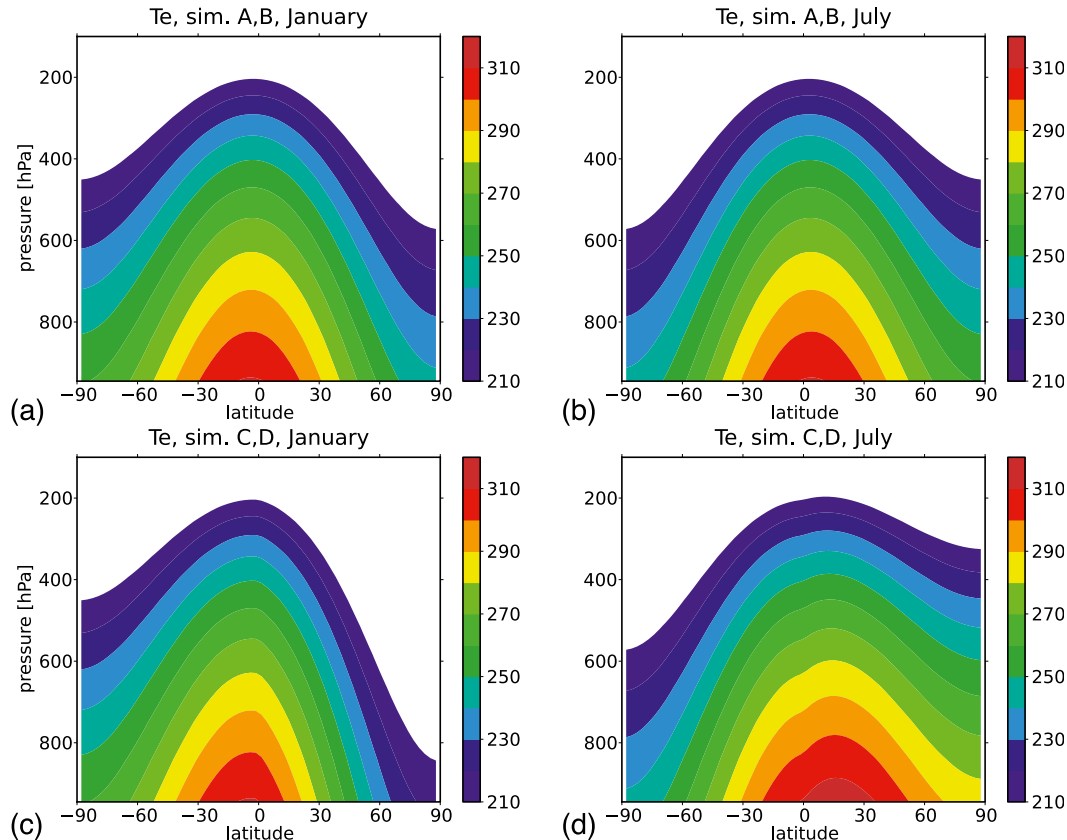


FIG. 8. Tropospheric T_e profiles for (a),(c) January and (b),(d) July. (top) The symmetric simulations HS94, PK02, GP09, A, and B. (bottom) Asymmetric simulations C and D. Notice the steeper meridional gradient in January and shallower gradient in July in the NH asymmetric case compared to the symmetric profiles.

between simulations B and C. For this, it is convenient to consider the results for averages over December–February (DJF) and June–August (JJA) separately. When comparing the results of the setups B and C, it is important to keep in mind that above about 200 hPa, there is no difference in the Newtonian cooling terms between these two cases. Figures 10a and 10b show the temperature differences between simulations B and C in the stratosphere. These differences are entirely due to differences in $\mathcal{T}$. In Fig. 10a, the dynamic temperature difference $T(C) - T(B)$ for DJF shows a cooling of the tropics, with a maximum around 70 hPa. The signature follows that of a stronger deep circulation, with increased cooling in the tropics (around 70 hPa), and warming in the extratropics. In contrast, we can derive from Fig. 10b that the layer from 70 to 100 hPa, including the cold-point tropopause, is substantially warmer for simulation C in JJA, which indicates a weaker shallow circulation. We conclude that the seasonal cycle in the TTL and the lower stratosphere is a combination of a stronger deep circulation in the northern winter hemisphere and a weaker shallow circulation in northern summer hemisphere.

We have found that the temperature of simulation C exhibits the structure of stronger (weaker) shallow residual circulation in NH winter (summer) when compared to simulation B. As the residual circulation is driven by eddy flux convergence, it is instructive to compare eddy fluxes between the two simulations and reanalysis. Figure 11 displays $\overline{v'T'}$ weighted by the cosine of latitude at 70 hPa for the two setups and at 66 hPa for ERA-Interim for January and July. We note that results for $\overline{u'v'}$ are very similar (not shown). In each of the three graphs, the July curve has been mirrored around the equator, such that the two winter sides are to the right and the two summer sides to the left of the equator, with the tropics highlighted with a somewhat darker background shading. The curve for January is in solid, the July curve is in dashed lines. Red shading between the two curves indicates stronger wave forcing in January and green shading indicates stronger forcing in July. We will first concentrate on Fig. 11a for the symmetric simulation B.

In the winter hemisphere, the January heat flux is generally larger, owing to the orographic forcing present

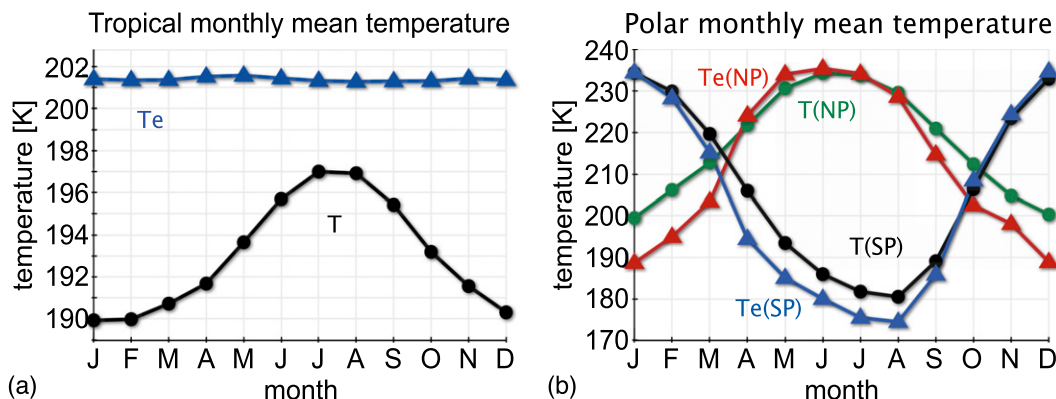


FIG. 9. Comparing dynamic to radiative effects for simulation C. (a) The seasonal cycles of T_e and T averaged from 20°S to 20°N at 96 hPa. (b) As in (a), but over the poles, averaged from 70° to 90° at 50 hPa in both hemispheres. Circles denote dynamic and triangles show relaxation temperatures.

in the NH only. In the summer hemisphere, the July heat flux is stronger, again owing to the presence of the mountains on the NH as the only asymmetry between NH and SH. In both hemispheres, the winter forcing is larger than the summer forcing, as only the polar vortex allows waves to propagate high into the stratosphere. Another way to describe the mid- to high-latitude results in simulation B is as follows: Winter hemisphere heat flux is stronger than summer hemisphere heat flux, and mountains increase heat flux compared to a flat surface. In the tropics, there is no difference between January and July. As a result, there is only little seasonal variation in tropical upwelling in the north–south symmetric tropospheric setup (Fig. 7d).

For the calculations with asymmetric setup in the troposphere (simulation C), and in observations (Figs. 11b,c), the results are rather different. There is no green shading, and the red shading penetrates deep into the

tropics, meaning that the January forcing is stronger than the July forcing everywhere, including in the tropics and the summer hemisphere. Thus, in the subtropics and summer hemisphere, baroclinic forcing is now more important than orographic forcing. Both the model and ERA-Interim show larger heat flux in January than in July (red shading) inside the tropics, although ERA-Interim is only on the winter side, whereas the model run has differences everywhere in the tropics. These plots suggest that our model reflects tropical sensitivities correctly in the sense that January wave drag has to be larger than July wave drag everywhere, not just the winter high latitudes.

b. Role of orographic forcing in asymmetric troposphere

The relationship between EP fluxes and dynamical variables such as zonal wind and temperatures is nonlinear.

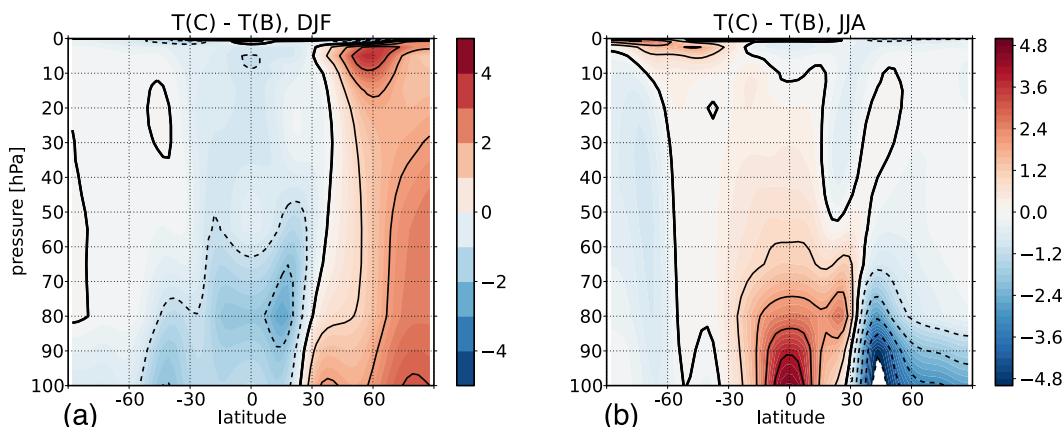


FIG. 10. Temperature differences between simulations C and B for (a) DJF and (b) JJA. Recall that the difference between the two simulations lies entirely in the setup of the troposphere (symmetric vs asymmetric) and there are no differences in T_e above 200 hPa.

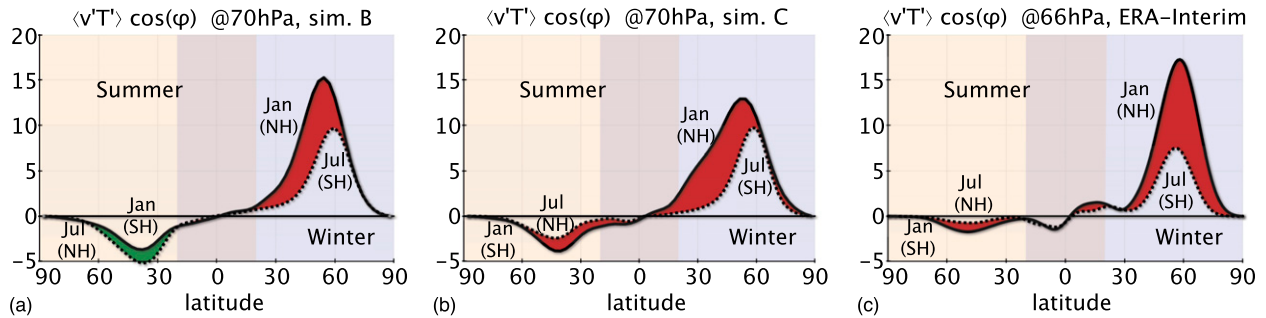


FIG. 11. $\overline{v'T'} \cos \phi$ at 70 hPa for (a) simulation B and (b) simulation C compared to (c) ERA-Interim at 66 hPa. The horizontal axis is divided into winter and summer hemispheres, with the July curves mirrored around the equator for direct comparison between winter and summer heat fluxes. Solid lines denote the January values, dashed lines denote the July values, and the labels in parentheses clarify which hemisphere is the winter hemisphere for the given month. Red shaded regions denote larger heat flux in January and green regions denote larger heat flux in July.

Therefore, it is difficult to attribute one specific observation (like the seasonal cycle of the TTL temperature) to one specific driving mechanism (as, e.g., $\partial_y T_e$). The simulations with our more realistic stratospheric setup, and our modifications to the HS94 troposphere, suggest that the topographic midlatitude forcing regulates mainly the polar vortex strength and the deep meridional circulation. In the TTL between 100 and 70 hPa, seasonal variation in tropospheric baroclinicity is the dominant contributor to temperature (and upwelling) variability on seasonal time scales. To check this, we ran the same setup as above but removed the orographic forcing, setting the mountain height to zero (simulation D). Figure 12 shows that, as expected, the NH polar vortex is colder with this setup (i.e., $|T|$ is smaller), with a minimum temperature comparable to the tropical cold point. Similarly, the amplitude of the seasonal temperature variation in the upper tropical and northern stratosphere is somewhat reduced, although not by a large factor. At the equator, a large-amplitude seasonal cycle

is still present in the TTL, owing to the unchanged seasonality in baroclinicity. However, the area of large variations in the tropics is much smaller now, and the maximum amplitude has been reduced by about a factor of 2.

Figure 13 plots the EP flux propagation together with zonal mean zonal wind for NH winter (top) and SH winter (bottom; mirrored around the equator in Figs. 13e and 13f for direct comparison). Remember that of these three columns, the left shows the weakest seasonal cycle in the tropics, the middle shows the strongest, and the right shows a temperature amplitude somewhat in between the other two. The differences between Figs. 13a and 13d are located entirely in the winter hemisphere, where more EP flux propagates into the stratosphere if orographic forcing is present (Fig. 13a). This is as expected, and the seasonal cycle of Fig. 4 can be attributed to orographic forcing. Notice how there is slightly stronger upper-tropospheric EP flux directed toward the tropics in the PK02 winter hemisphere (Fig. 13d). Even

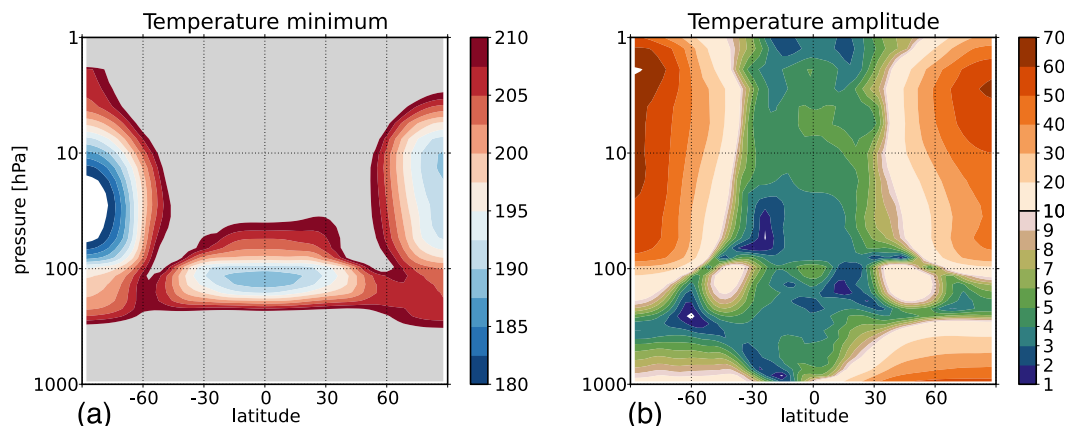


FIG. 12. Temperature (a) minimum and (b) amplitude with our new stratospheric setup, as in Fig. 7, but for simulation D. The only difference with simulation C (Figs. 7e,f) is the absence of topography.

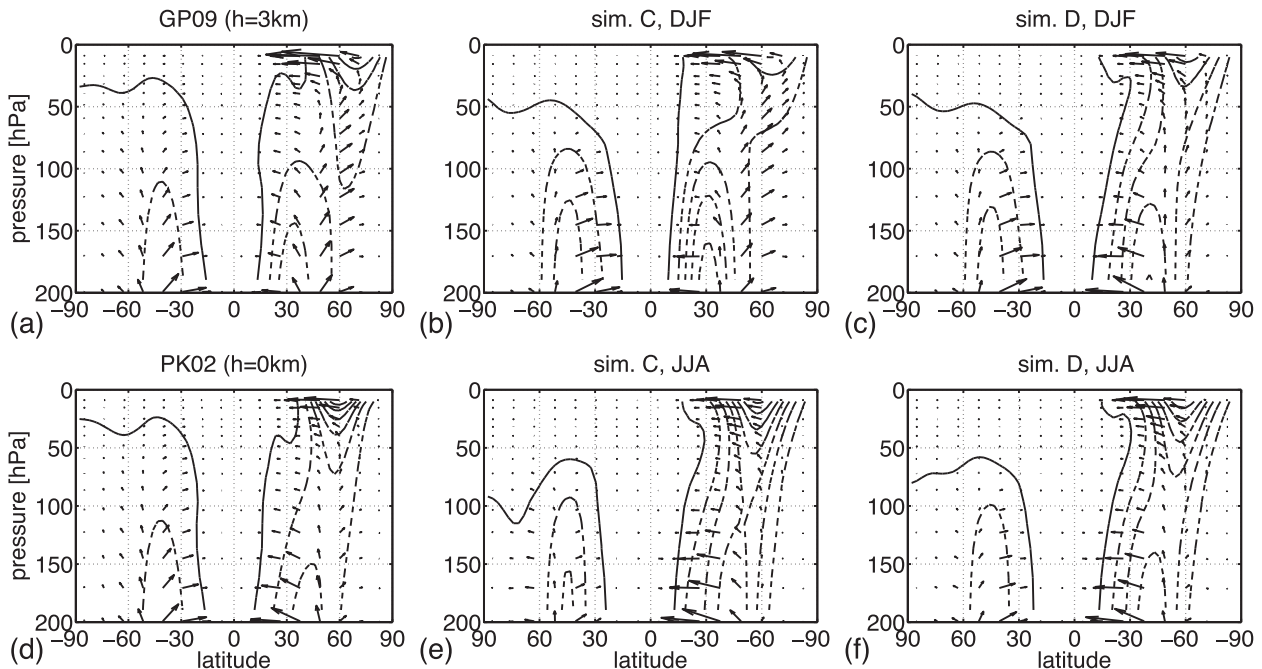


FIG. 13. EP fluxes (arrows) and zonal wind (contours) for (a),(d) the GP09–PK02 setup, (b),(e) simulation C (with mountain), and (c),(f) simulation D (without mountain). For easier direct comparison with the DJF results in (b),(c), we have inverted the meridional axis for the JJA figures in (e),(f). Zonal wind contour interval is 10 m s^{-1} , and the solid line denotes the zero wind line. (middle),(right) Arrows are scaled identically, and (left) arrows are doubled.

though it is a second-order effect, this additional upper-tropospheric EP flux converging in the tropics could further diminish the seasonal-cycle amplitude, as it compensates for a (small) part of the missing orographic forcing in upwelling. For simulations C and D, there are rather strong differences in the summer hemispheres (cf. Fig. 13b to Fig. 13e and Fig. 13c to Fig. 13f), in addition to the differences due to topography in the winter hemispheres. The strong seasonal cycle in the tropics for simulation C comes not only from larger midlatitude EP fluxes in DJF than in JJA but also from the differences in the summer hemispheres. This is why even though simulation D does not have any topography, it still has a stronger lower-stratospheric tropical seasonal cycle than the GP09 and PK02 simulations.

Thus, even with a more realistic stratospheric setup, the state of the troposphere is an important factor for the emergence of a seasonal cycle in lower-stratospheric tropical upwelling: In our model, the observed seesaw in tropical versus polar temperatures arises from a combination of variations in orographic forcing and baroclinic eddy activity in the troposphere. It is difficult to make a strong quantitative statement as to how important baroclinic eddies are with respect to orographic forcing. To do so, we will have to conduct more rigorous parameter space studies, as done for example by Gerber (2012), but this is left for future work.

6. Summary and conclusions

We have analyzed the structure and variability of the lower stratosphere in dry GCM calculations based on widely used Newtonian cooling setups. We find that these setups do not give a particularly realistic lower stratosphere, with deficiencies particularly pronounced in the tropics. Based on radiative transfer calculations, we derived a new Newtonian cooling setup that improves the lower stratosphere substantially. In particular, the locations of the tropical tropopause and subtropical jets are improved, and the temperature minima at the tropical tropopause (the tropical “cold point”) and the northern and southern polar vortices are more realistic. The radiative equilibrium temperature field that we propose is a numerical field, but in future work we will present an analytic approximation to it.

Using various forms of thermodynamic forcing in conjunction with a wave-2 surface topography in one hemisphere, we have evaluated the impact of hemispheric asymmetric quasi-stationary planetary wave forcing on the stratospheric circulation, keeping the thermodynamic forcing of the troposphere fixed and hemispherically symmetric. As perhaps expected, our results show that this midlatitude wave-2 forcing substantially affects the mid- and high-latitude stratospheric circulation. Its impact on the tropics, however, is too small to explain

the observed amplitude of lower-stratospheric seasonal temperature variations, even when using large values of topography, for all types of stratospheric forcing. Thus, for example, in the GP09 setup, using a wave-2 forcing during winter in one hemisphere gives a seasonal temperature variation of only about 1 K, whereas observations show a variation of about 8 K. The radiatively determined setup introduced here gives a dynamically forced amplitude of about 3 K, mostly because of a more realistic zonal wind structure, which allows for a better representation of wave propagation into the stratosphere, but the amplitude is still too small. In agreement with previous studies, we find that including the effect of seasonal variations of ozone increases the amplitude by about 1 K to a total of 4 K—still too small. We conclude that a seasonal cycle in stationary wave forcing alone is insufficient to produce a tropical stratospheric seasonal cycle of sufficient amplitude.

Our simulations do, of course, explicitly produce baroclinic instability that, despite being at relatively large wavenumbers, produces waves that can, depending on the zonal wind structure, propagate into the lower stratosphere where they dissipate and force a lower branch of the stratospheric residual circulation. If we allow baroclinic instability to vary seasonally and hemispherically, by changing the thermodynamic forcing in the troposphere, we are able to produce large variations in tropical upwelling. Indeed, adding such tropospheric effects to the newly derived stratospheric structure increases the amplitude of the seasonal cycle in the tropical lower-stratospheric temperature to 8 K—in good agreement with observations.

Thus, to summarize, we are able to reproduce the observed tropical lower-stratospheric temperature seasonal cycle with an idealized model provided that all of the following three factors are present: (i) a reasonably realistic basic stratospheric structure determined by the local thermodynamic forcing, (ii) seasonally varying topographically forced planetary waves, and (iii) seasonally varying baroclinic instability in the troposphere. The amplitude of the cycle is about 8 K and this can be roughly attributed as follows. About 4 K comes from hemispheric asymmetries in tropospheric meridional temperature gradient, with a stronger seasonal cycle in the Northern Hemisphere; about 3 K comes from extratropical planetary wave driving, provided the stratospheric zonal wind structure is realistic; and finally, a small part (about 1 K) comes from stratospheric tracer variations: in particular, ozone. Although the partitioning of these factors is not exact and a comprehensive GCM might yield different results, we believe that all of the above-mentioned processes are important to some degree.

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