Traveling planetary-scale Rossby waves in the winter stratosphere: The role of tropospheric baroclinic instability

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[1] The Southern Hemisphere winter stratosphere exhibits prominent traveling planetary-scale Rossby waves, which generally are not able to induce Stratospheric Sudden Warmings. A series of runs of a simplified general circulation model is presented, aimed at better understanding the generation of these waves. While the generation of planetary-scale traveling waves through the interaction of synoptic-scale waves is observed in a control run, when the model is truncated to permit only waves with zonal wave number 1 or 2, the long waves are found to increase in strength, leading to a considerably more active stratosphere including Sudden Warmings comparable in strength to Northern Hemisphere winter. This finding suggests that the role of tropospheric synoptic eddies is two-fold: while generating a weak planetary-scale wave flux into the stratosphere, their main effect is to suppress baroclinic instability of planetary-scale waves by stabilizing the tropospheric mean state. **Citation:** Domeisen, D. I. V., and R. A. Plumb (2012), Traveling planetary-scale Rossby waves in the winter stratosphere: The role of tropospheric baroclinic instability, *Geophys. Res. Lett.*, *39*, L20817, doi:10.1029/2012GL053684.

1. Introduction

[2] Traveling planetary-scale Rossby waves are observed in the extratropical stratosphere of both hemispheres. They are especially prominent in the Southern Hemisphere, while quasi-stationary waves dominate in the Northern Hemisphere due to the presence of stronger surface excitation. Several mechanisms have been proposed to be able to force large amplitude traveling waves in the absence of strong surface forcing. *Hartmann* [1979], for instance, suggested a generation of traveling waves by Charney-type baroclinic instability of zonal wave-2 in the troposphere. In contrast, *Scinocca and Haynes* [1998] identified nonlinear interactions among synoptic-scale baroclinic eddies in the troposphere as the origin of stratospheric traveling waves. We here report results from a dynamical core model integration without external wave forcing. It is in particular shown that suppressing synoptic-scale eddies by a severe spectral truncation yields an increase in planetary wave flux into the stratosphere. While the absence of synoptic eddies inhibits the excitation of planetary-scale motion according to the

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Scinocca and Haynes [1998] mechanism, it also changes the mean state to permit long-wave generation via baroclinic instability [*Hartmann*, 1979] in the troposphere.

2. Model Setup

[3] The model used in this study is the spectral core of a general circulation model at T42 resolution with 40 vertical levels as specified in *Polvani and Kushner* [2002], except the present model is in hybrid $\sigma - p$ coordinates. This setup includes a linear relaxation towards a zonally symmetric equilibrium temperature profile which corresponds to the *Held and Suarez* [1994] profile in the troposphere with an asymmetry about the winter hemisphere, and a cooling over the winter pole in the upper stratosphere. We use a $\gamma =$ 4 K/km lapse rate for the winter stratospheric cooling (for the definition and use of γ, see *Polvani and Kushner* [2002]), which corresponds to a strong Southern Hemisphere- like polar vortex. Each model run is integrated for 10,000 days and the last 9800 days are used in the analysis.

[4] We compare the control run to two truncated runs: For the truncated runs, the only difference to the control run is a truncation in wave number space to zonal wave-1, wave-2, and a mean flow only (hereafter: *trunc12*) and zonal wave-2 and a mean flow (hereafter: *trunc2*), respectively. There is no such truncation in the meridional direction.

3. Results

[5] The control run yields a very strong polar vortex with mean winds around 100 ms^{-1} at its center, with a standard deviation of 5 ms^{-1} (Figures 1a and 2a). Stratospheric variability is significantly reduced as compared to the real atmosphere; in particular, no stratospheric sudden warmings are observed. Remarkably, the truncated runs show much more variability (Figures 1b and 1c), including intermittent large-amplitude warming events, with a corresponding reduction in the time-mean strength of the stratospheric jet (Figures 2b and 2c).

[6] The changes in the stratospheric long waves that accompany these changes in the model behavior are relatively modest. Figure 3 shows the time-averaged vertical component of the Eliassen-Palm flux for zonal waves 1 and 2 in the three runs. The truncated runs show fluxes that are much larger in the troposphere, but just weakly increased in the stratosphere. The characteristics of the waves are affected by the truncation, however. Since there is no external forcing in these experiments, quasi-stationary waves are essentially absent. In both the control and the truncated runs zonal wave-2 exhibits intermittent occurrences of systematic eastward propagation with periods of around 10 days interspersed with episodes of slower propagation in either direction. Figure 4 shows phase speed spectra computed from geopotential height amplitude

a) zonal mean zonal wind, control, 10hPa, 60S

b) zonal mean zonal wind, trunc12, 10hPa, 60S

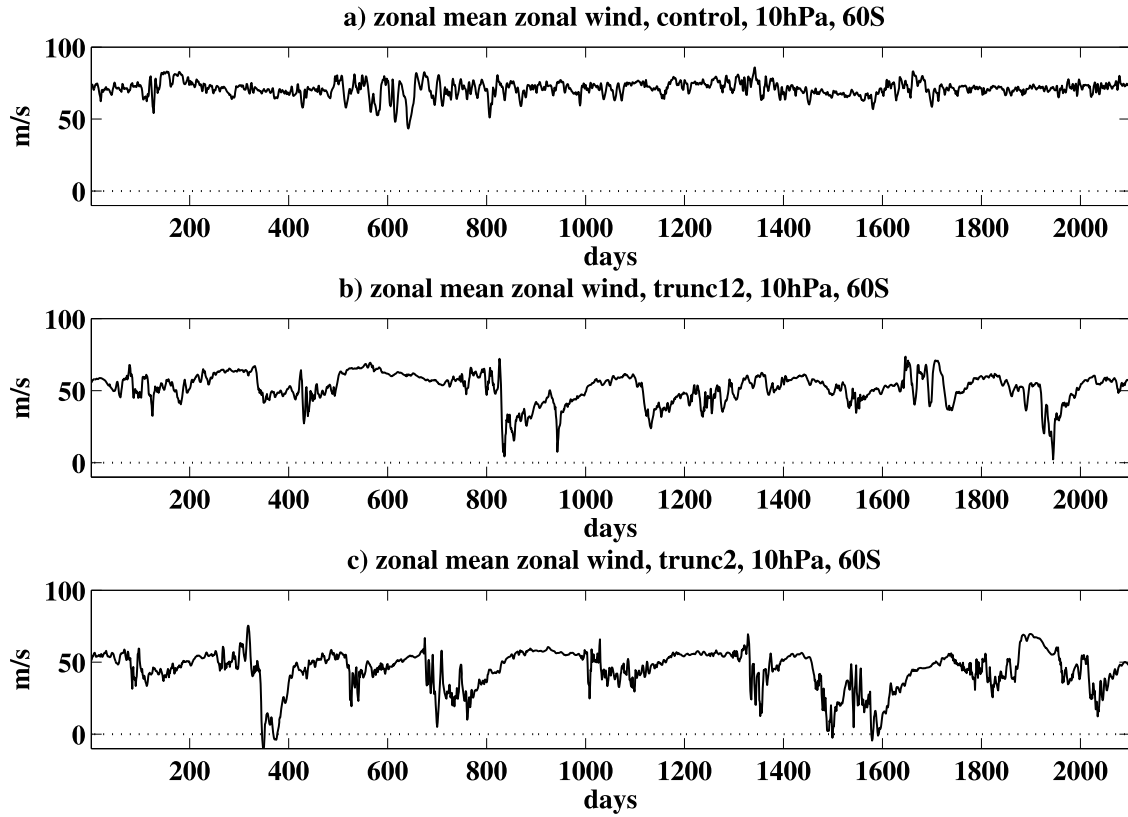### c) zonal mean zonal wind, trunc2, 10hPa, 60S

Figure 1. Representative part of the time series of zonal mean zonal wind at 60°S and 10 hPa for (a) the control run, (b) the *trunc12* run, and (c) the *trunc2* run.

of wave-2 at 189 hPa. In the control run there is broad low frequency variability, while in both truncated runs, there is a more dominant peak around phase speeds of 18 ms^{-1} eastward.

[7] Since in the truncated runs, motions of zonally synoptic scale are eliminated, nonlinear interaction between them cannot be responsible for forcing the long waves. Tropospheric baroclinic instability of the long waves themselves,
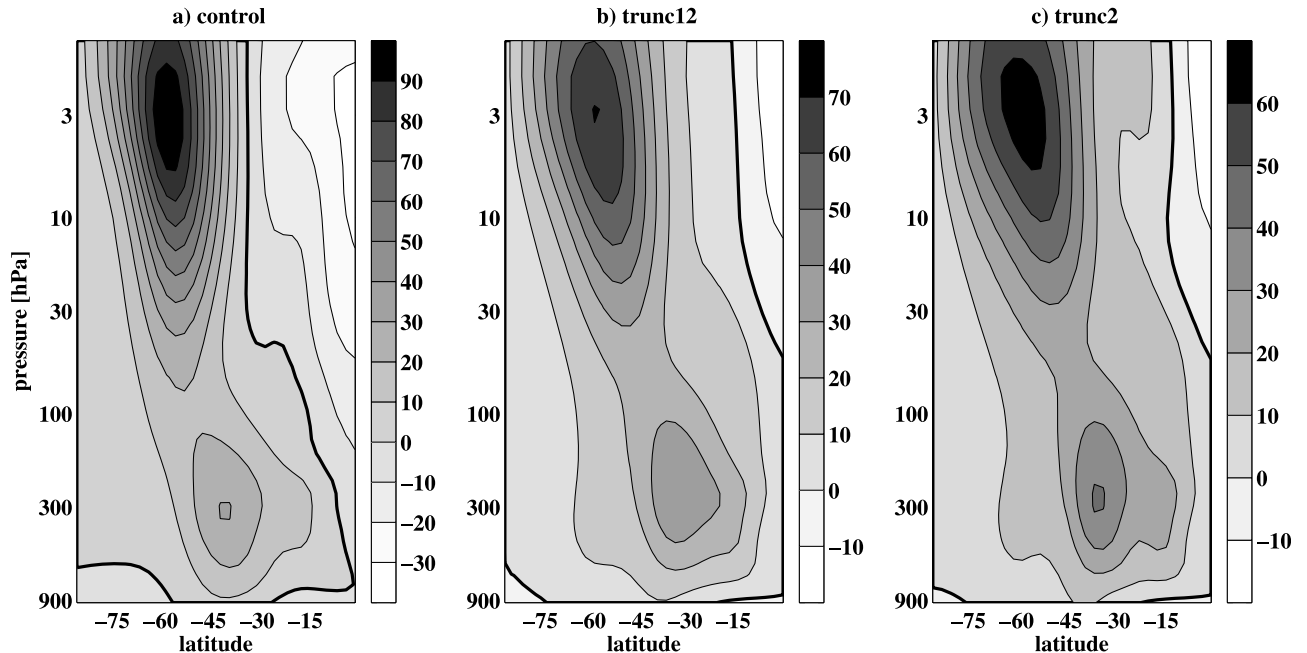
a) control

b) trunc12

c) trunc2

Figure 2. Time averaged zonal mean zonal wind for (a) the control run, (b) the *trunc12* run, and (c) the *trunc2* run. Contour interval: 10 ms^{-1}. Zero wind line printed in bold.
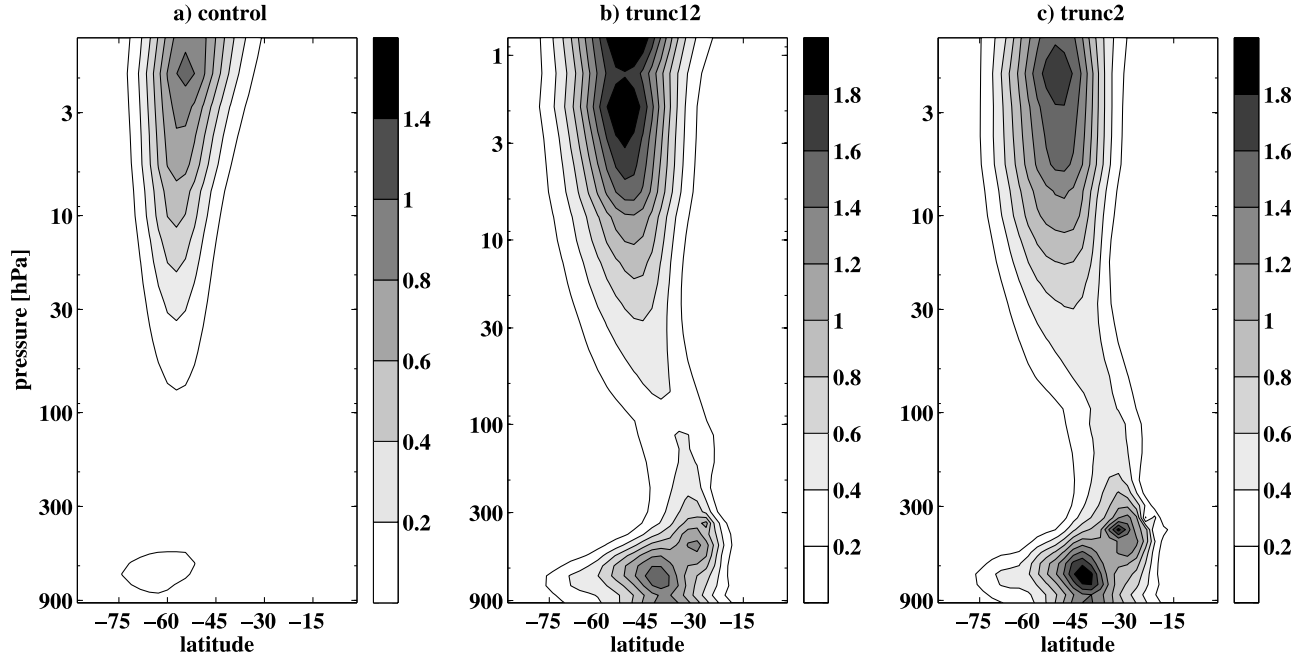
Figure 3. Vertical component of the EP flux scaled by density $[F_z/\rho_0(z)]$, sum of wave-1 and 2 for (a) the control run, (b) the *trunc12* run, and (c) the *trunc2* run. Units: 10^6 m^3 s^{-2}. Contour interval: $2 \cdot 10^5$ m^3 s^{-2}. Contours start at $2 \cdot 10^5$ m^3 s^{-2}, zero and negative contours are omitted for clarity.

as described by *Hartmann* [1979], is therefore suggested to be responsible for the generation of these waves. *Hartmann* [1979] identified three main requirements for the occurrence of baroclinic instability of planetary-scale waves: an increased vertical wind shear, wave growth at high latitudes, and a small meridional length scale of the wave.

[8] Increased vertical wind shear and increased lower tropospheric baroclinicity in the truncated runs is evident in the zonal wind plots of Figure 2, a difference that can be ascribed to the stabilizing influence of the synoptic-scale eddies in the control run. We quantify this by showing in Figure 5 the Eady growth rate

$$\sigma = \left| \frac{f_0}{N} \frac{\partial \overline{u}}{\partial z} \right| \qquad (1)$$

as a measure of the growth rate of the most unstable mode in the model atmosphere, where N is the buoyancy frequency. The derivatives are evaluated in the lower troposphere between the model levels at 514 hPa and 925 hPa. Figure 6 shows a comparison between the meridional length scale of zonal wave 2 for the presented runs. The variable shown is the Southern Hemisphere meridional length scale L_y of the wave scaled by Earth's radius a, defined here, assuming geostrophic balance, as

$$\frac{L_y}{a} = -\frac{g}{f\,a}\sqrt{\frac{\left\langle \overline{\phi_2'^2} \right\rangle}{\left\langle \overline{u_2'^2} \right\rangle}}, \qquad (2)$$

where f is the Coriolis parameter, and $\left\langle \overline{\phi_2'^2} \right\rangle$ and $\left\langle \overline{u_2'^2} \right\rangle$ are the time averaged zonal variances of wave-2 geopotential and zonal wind. Both the control run and the truncated runs indicate a small meridional scale, $L_y \leq 0.2\,a$, of the wave in

the troposphere, with the scale being smallest in the truncated runs. The scale increases with height as required for propagation into the stratosphere [*Charney and Drazin*, 1961].

4. Discussion and Conclusions

[9] In addition to the forcing of quasi-stationary planetary-scale waves by topography or zonally asymmetric tropospheric heat sources, the generation of transient, zonally long waves penetrating the winter stratosphere is, following *Scinocca and Haynes* [1998], attributed to nonlinear wave-wave interaction between synoptic-scale baroclinic eddies in the troposphere, rather than being a direct consequence of baroclinic instability as suggested by *Hartmann* [1979]. In the absence of external, stationary, forcings such as in numerical experiments like those described here, transient long waves thus generated may propagate into the winter stratosphere, though in such circumstances they are generally not able to produce much stratospheric variability.

[10] The strengthening of these long waves in the truncated runs is so severe as to generate major warmings within the model stratosphere, an effect that requires large amplitude surface topography in the control run [*Gerber and Polvani*, 2009]. This dramatic impact of the removal of synoptic-scale eddies in the truncated runs reveals that the role of the synoptic eddies in the control case and, by implication, in the atmosphere, is two-fold. In addition to providing, through their mutual interactions, a source of transient long waves as described by *Scinocca and Haynes* [1998], they also act to stabilize the troposphere and hence to suppress the generation of long waves via baroclinic instability. Results from the control run (and from prior work, especially that of *Scinocca and Haynes* [1998]) imply that in realistic climates this suppression is essentially total, leaving the *Scinocca and Haynes* [1998] process as the
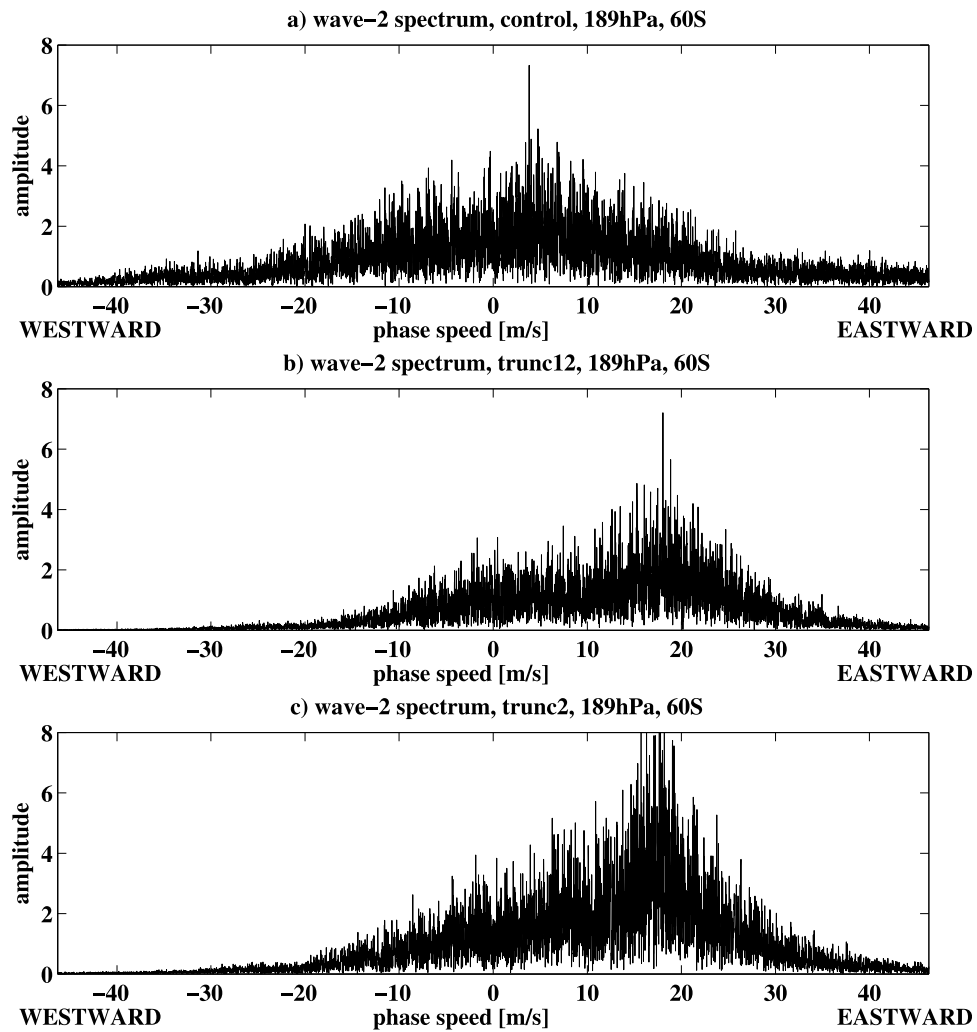
Figure 4. Phase speed spectra computed from geopotential height for zonal wave-2 at 189 hPa and 60°S for (a) the control run, (b) the *trunc12* run, and (c) the *trunc2* run.
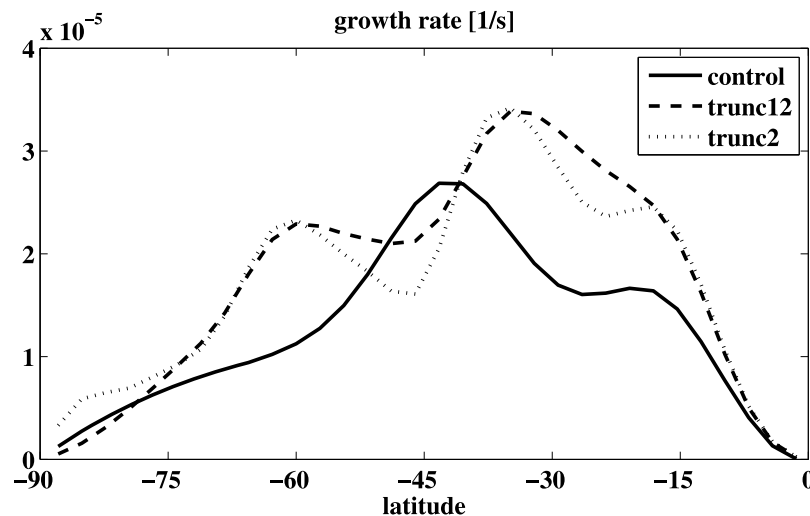

Figure 5. Comparison of the growth rates σ [s^{-1}] as a function of latitude (equation (1)), proportional to the Eady growth rate for the most unstable mode.
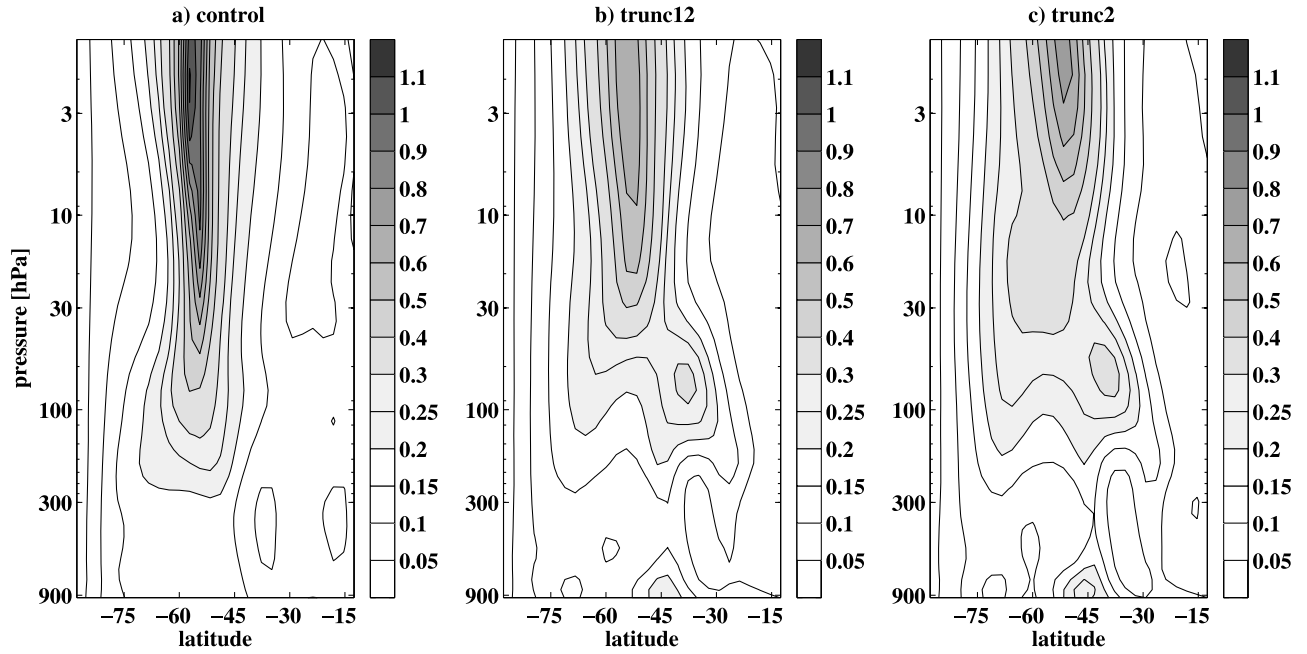
Figure 6. L_y/a (equation (2)) as a measure of the meridional length scale of wave-2 for (a) the control run, (b) the *trunc12* run, and (c) the *trunc2* run. Contour interval: 0.1.

dominant forcing mechanism, especially in the Southern Hemisphere, where topographic forcing is relatively weak.

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