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

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We report results from a series of runs of a simplified (“dynamical core”) general circulation model, aimed at understanding the generation of traveling planetary-scale Rossby waves in the winter stratosphere. The model has no topographic or other external wave forcing. When the model is truncated to permit only waves with zonal wave number 1 or 2, the long waves are found to be stronger than when zonally shorter waves are included, leading to a more active stratosphere. On the basis of these results and the diagnosed wave properties, we conclude that (consistent with an earlier suggestion by Hartmann) traveling long waves are generated primarily by tropospheric baroclinic instability rather than by nonlinear interactions among synoptic-scale eddies.
1. Introduction

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] suggested a generation of traveling waves by Charney-type baroclinic instability of zonal wave-2 in the troposphere. Scinocca and Haynes [1998] suggested nonlinear interactions among synoptic-scale baroclinic eddies in the troposphere as the origin of stratospheric traveling waves. Kushner and Polvani [2005] ascribed waves (and a major stratospheric warming produced by these waves) in a spectral core model with no longitudinally varying forcing to the latter mechanism.

We here report results from a similar spectral core model without external wave forcing: Suppressing synoptic-scale eddies by a severe spectral truncation yields an increase in planetary wave fluxes into the stratosphere. The absence of synoptic eddies necessary to excite planetary-scale motion according to the Scinocca and Haynes [1998] mechanism, together with the characteristics of the modeled wave structure, leads us to conclude that the model’s planetary-scale waves are produced primarily by Charney-type baroclinic instability in the troposphere, as Hartmann suggested.

2. Model Setup

The model used in this study is the spectral core of a general circulation model of intermediate complexity at T42 resolution 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 zonal mean 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 $\gamma=4K/km$ lapse rate for the winter stratospheric cooling (for the definition and use of $\gamma$ see Polvani and Kushner [2002]), which corresponds to a strong Southern Hemisphere-like polar vortex. The model runs have no seasonal cycle and are run in perpetual winter conditions.

We are presenting a comparison of two model runs of 10,000 days length each: Both runs use the above setup with no longitudinally varying forcing. We then compare the control run to a truncated run as explained below.

For the truncated run, the only difference to the control run is a truncation in wave number space to zonal wave-1 and wave-2 and a mean flow only. There is no such truncation in the meridional direction.

For completeness, an additional model run was performed which was truncated to zonal wave-2 and a mean flow only. It confirms the conclusions made from the above runs as described in the Results section.

3. Results

The control run yields a very strong polar vortex with mean winds around 100m/s at its center, with a standard deviation of 5m/s. Stratospheric variability is significantly reduced as compared to the real atmosphere, and no stratospheric sudden warmings are
observed (Figure 1a shows a representative excerpt from the control run). The time series of the Eliassen-Palm flux (EP flux) entering the stratosphere looks noisy (Figure 1b).

Truncating the model to zonal wave-1 and wave-2 yields a significant increase in stratospheric variability (Figure 2a). In particular, large amplitude warmings occur intermittently associated with large excursions in the EP fluxes (Figure 2b).

In order to understand the differences in wave behavior, time-averaged vertical EP fluxes divided into planetary-scale (zonal wave-1 and wave-2) and shorter (wave-3 and higher) waves are examined. In the control run the long waves determine the wave flux in the stratosphere as expected (Figure 3b), while the EP flux in the troposphere is dominated by higher wave numbers (Figure 3c), with the long waves accounting for only about 10% of the tropospheric wave flux.

If these shorter waves are responsible for the upward long-wave EP flux in the stratosphere, we would expect the long-wave flux to vanish when smaller scale waves are inhibited in the model atmosphere. In the truncated run, however, the long-wave flux actually strengthens, not only in the stratosphere, but also in the troposphere (Figure 4b). Results from the additional model run truncated to zonal wave-2 and a mean flow yield the same results, in fact this run exhibits slightly stronger stratospheric variability than the run truncated to both wave-1 and wave-2.

Since in the truncated run, synoptic motions are nonexistent, nonlinear interaction between them cannot be responsible for forcing the long waves. Tropospheric baroclinic instability of the long waves themselves, as suggested by Hartmann [1979], must be responsible for the generation of these waves.
Instability of the long waves is enhanced in the truncated run since the suppression of
the synoptic-scale waves increases the baroclinicity of the troposphere (compare the lower
tropospheric wind shear in Figures 3a and 4a).

In addition, the characteristics of the waves indicate their origin from baroclinic insta-
bility. Waves with long zonal wavelengths can become synoptic-scale in terms of their
total horizontal wavelength by adopting a large meridional wavenumber, and by being
confined to high latitudes [Hartmann, 1979], and can be classified as Charney modes.

Figure 5 shows a comparison between the meridional length scale of zonal wave-2 for the
two 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

$$\frac{L_y}{a} = -\frac{g}{f a} \sqrt{\frac{<\phi_2'^2>}{<u_2'^2>}}$$

where $f$ is the Coriolis parameter, $\phi_2'$ are the geopotential wave-2 zonal anomalies, $u_2'$ are
the wave-2 zonal anomalies of zonal wind, an overbar denotes a zonal average and $<.>$
denotes a time average. Both the control run and the truncated run indicate a small
meridional scale, $L_y \leq 0.2a$, of the wave in the troposphere. The scale increases with
height as required for propagation [Charney and Drazin, 1961].

Since there is no external forcing in these experiments, quasi-stationary waves are es-
sentially absent. In both the control and the truncated runs zonal wave-2 exhibits inter-
mittent occurrences of systematic eastward propagation with periods of order 10 days
interspersed with episodes of slower propagation in either direction. Figure 6 shows spec-
tra of geopotential height amplitude of wave-2 at 189hPa. There are some differences between the characteristics of the control and the truncated run: In the latter, there is a spectral peak of about frequency 0.08/day, whereas in the control run there is broad low frequency variability without a clear peak.

4. Conclusions

In the absence of a stationary planetary-scale tropospheric forcing such as topography or a heat source, the generation of zonally long waves penetrating the stratosphere has been attributed to nonlinear wave-wave interaction between synoptic-scale baroclinic eddies in the upper troposphere [Scinocca and Haynes, 1998].

However, stratospheric variability in a spectral core model with no longitudinally varying forcing increases when synoptic variability is eliminated, indicating that another mechanism, namely baroclinic instability of long waves, is responsible for forcing the long waves propagating into the stratosphere, as suggested by Hartmann [1979]. This hypothesis is supported by multiple indicators: wave-wave interaction among synoptic-scale waves is inhibited and therefore not responsible for causing the planetary wave flux into the stratosphere, interaction between planetary-scale waves weakens the wave flux into the stratosphere as confirmed in an additional experiment truncated to zonal wave-2 only, and the small meridional scale of the waves in the troposphere indicates that these modes are produced by tropospheric baroclinic instability.

Although we cannot rule out the possibility of a contribution from synoptic wave-wave interaction in the control run, the similarity of long-wave characteristics in the control and
truncated runs suggests that tropospheric baroclinic instability is the dominant long-wave
generation mechanism in both cases.

These results reinforce Hartmann’s suggestion that traveling planetary waves observed
especially in the Southern Hemisphere stratosphere may be the product of tropospheric
baroclinic instability rather than of nonlinear interaction between synoptic-scale eddies.

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References

Charney, J. and P. Drazin, 1961: Propagation of planetary-scale disturbances from the

Hartmann, D., 1979: Baroclinic instability of realistic zonal-mean states to planetary

Held, I. and M. Suarez, 1994: A proposal for the intercomparison of the dynamical cores

Kushner, P. and L. M. Polvani, 2005: A very large, spontaneous stratospheric sudden
warming in a simple AGCM: A prototype for the southern hemisphere warming of

Polvani, L. and P. Kushner, 2002: Tropospheric response to stratospheric perturbations

Scinocca, J. F. and P. Haynes, 1998: Dynamical forcing of stratospheric planetary waves
Figure 1. Control run: a) Representative part of the timeseries of zonal mean zonal wind at 60°S and 10hPa in m/s. b) Zonally averaged heat flux for wave-1 and wave-2 at 96hPa for the same time period, integrated between 20° and 70°S according to $\int_{20^\circ S}^{70^\circ S} F_z \cos(\theta) a \, d\theta$ where $F_z$ is the vertical component of the Eliassen-Palm flux in spherical coordinates, $\theta$ is latitude and $a$ is the Earth's radius.
Figure 2. Truncated run: a) Representative part of the timeseries of zonal mean zonal wind at 60°S and 10hPa in m/s. b) Zonally averaged heat flux for wave-1 and wave-2 at 96hPa integrated between 20° and 70°S as for the control run, computed as in Figure 1b.
Figure 3. Control run: a) Zonal mean zonal wind averaged over the entire run. Contour interval: 10m/s. Zero wind line printed in bold. b) Vertical component of the Eliassen-Palm Flux ($F_z$) scaled with density, sum of both propagating waves 1 and 2. Units in $10^6 m^3 s^{-2}$. Contour interval: $2 \cdot 10^5 m^3 s^{-2}$ with contours starting at $2 \cdot 10^5 m^3 s^{-2}$. Zero and negative contours omitted for clarity. c) Same as b) but for wave numbers 3 and higher.
Figure 4. Truncated run (wave-1 and wave-2 only): a) Zonal mean zonal wind averaged over the entire run. Contour interval: 10 m/s. Zero wind line printed in bold. b) Vertical component of the Eliassen-Palm Flux ($F_z$) scaled with density, sum of both propagating waves 1 and 2. Units and contours as in Figure 3.
Figure 5. a) $L_y/a$ for zonal wave-2 as a measure of the meridional length scale of the wave in latitude and height as described in equation (1) for the control run. b) Same as figure a) but for the run truncated to wave-1 and wave-2. Contour interval: 0.1 for both figures.
Figure 6.  a) Frequency spectrum of geopotential height amplitude for zonal wave-2 at 189hPa, averaged in longitude, for the control run.  b) Same as figure a) but for the run truncated to wave-1 and wave-2.