Stratospheric Response to Latitudinally Varied Surface Warming

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Introduction

Surface temperatures of a doubled-CO_2 world can be approximated by horizontal bands of warm anomalies, the strongest of which occur in the northern mid- to high latitudes, due to the land masses there. We have used an atmospheric GCM to simulate the effects of applying idealized, zonally symmetric positive temperature forcings at the surface in distinct latitude bands. We study both the response of the stratospheric circulation, and the sensitivity of this response to magnitude and location of the applied forcing.

In the control run, monthly mean surface temperatures and surface specific humidities are prescribed; these come from a separate 100-year run in which the model interacted freely with a slab ocean and a soil scheme.

In the experiments, a surface temperature forcing of 2 K is applied in the following latitude bands:
60°S – 30°S (case SM2), 30°S – 30°N (case EQ2), and 30°N – 60°N (case NM2).

Prescribed surface specific humidity is consistent with the temperature forcing.

We use the IGCM-FASTOC at T31 resolution with 26 vertical levels [1], [2], [3] and run it in timeslice mode for 50 years for each experiment. Results shown are the ensemble-mean values.

All plots show differences between the three experiments and the control run. Grey shading on all plots covers areas in which results are NOT significant at the 5% level, and green lines on all plots show where surface forcing is applied.

Temperature response

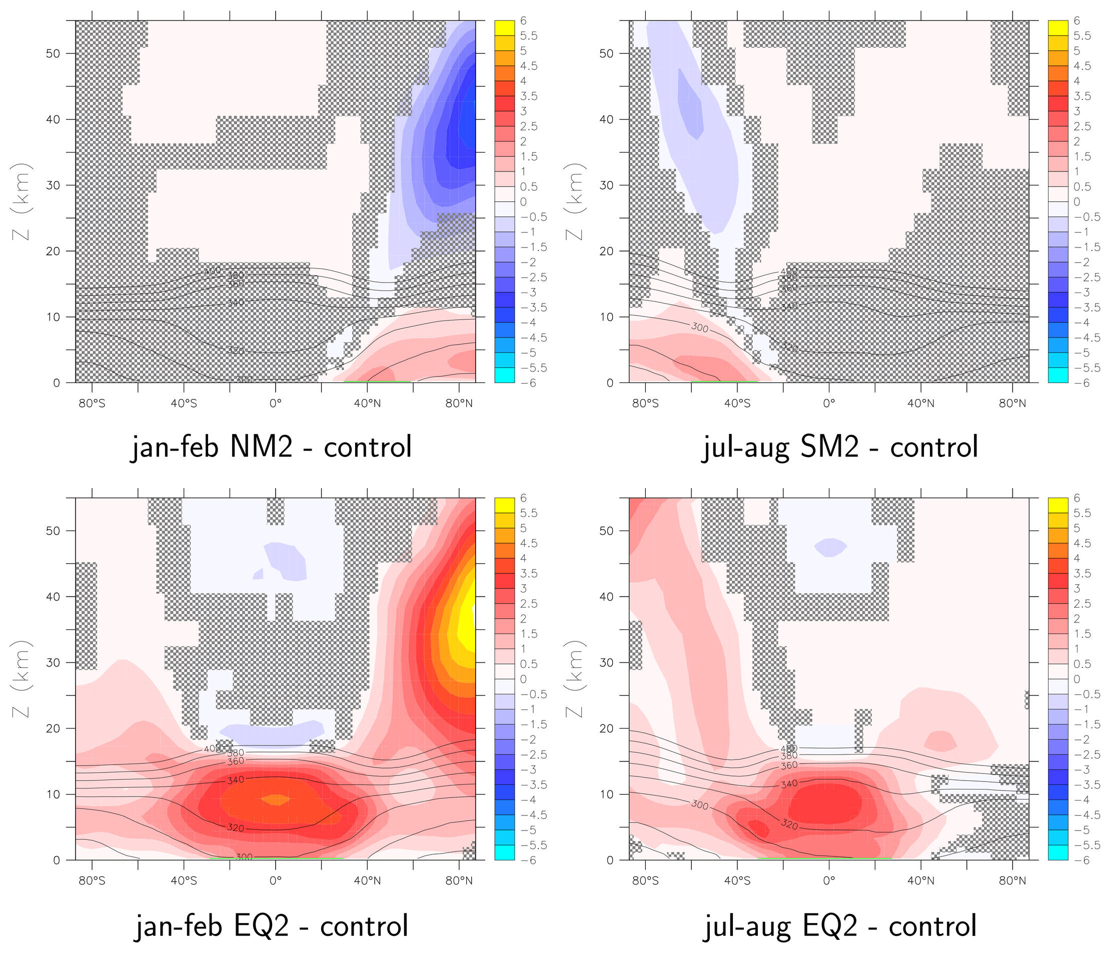

Fig 1: Zonal-mean temperature difference (K), with contours of control potential temperature (K).

- Top two panels: In each winter hemisphere, imposed mid-latitude surface temperature forcing is seen as a warm anomaly in the lower troposphere, but gives rise to a cold anomaly in the polar winter stratosphere. Responses to mid-latitude forcings are qualitatively the same in both hemispheres, though weaker in the SH.
- Bottom two panels: Surface temperature forcing in the EQ band warms the entire troposphere above the forcing area by more than the imposed 2 degrees. A strong warm anomaly also ensues in the NH winter polar stratosphere, whose SH counterpart is much weaker.
- The wintertime response in the two hemispheres is not symmetric, despite identical surface forcing.

Zonal wind response

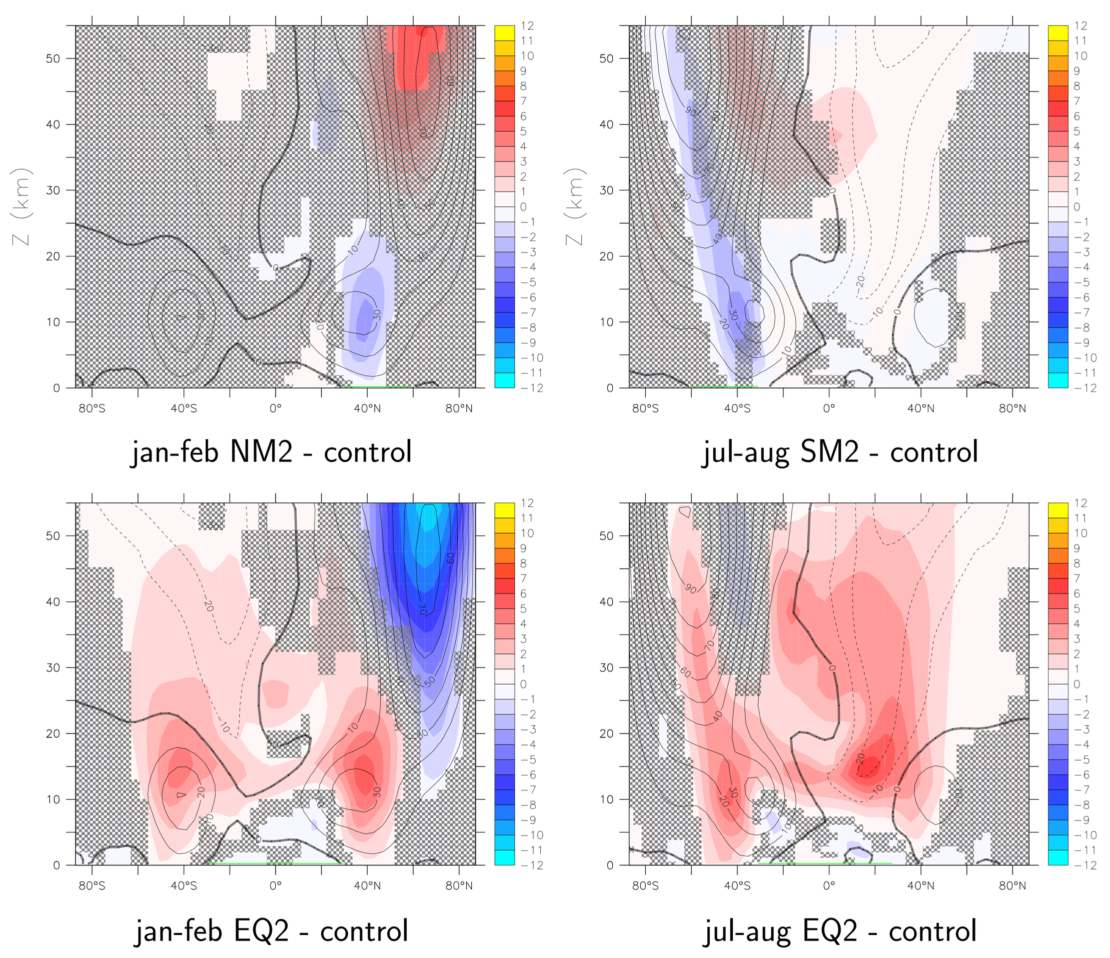

Fig 2: Zonal wind difference (m/s), with contours of control zonal wind.

- Top two panels: in accordance with thermal wind balance, the warm surface anomalies at high latitudes lead to weaker tropospheric jets. The cold NH anomaly in the polar stratosphere results in a slightly stronger polar vortex.
- Bottom two panels: the large increase in temperature throughout the tropics enhances the meridional temperature gradient and leads to stronger tropospheric jets in both hemispheres. The strong warm anomaly in the NH polar stratosphere is associated with a weaker polar vortex.

Wave forcing (NM2, EQ2)

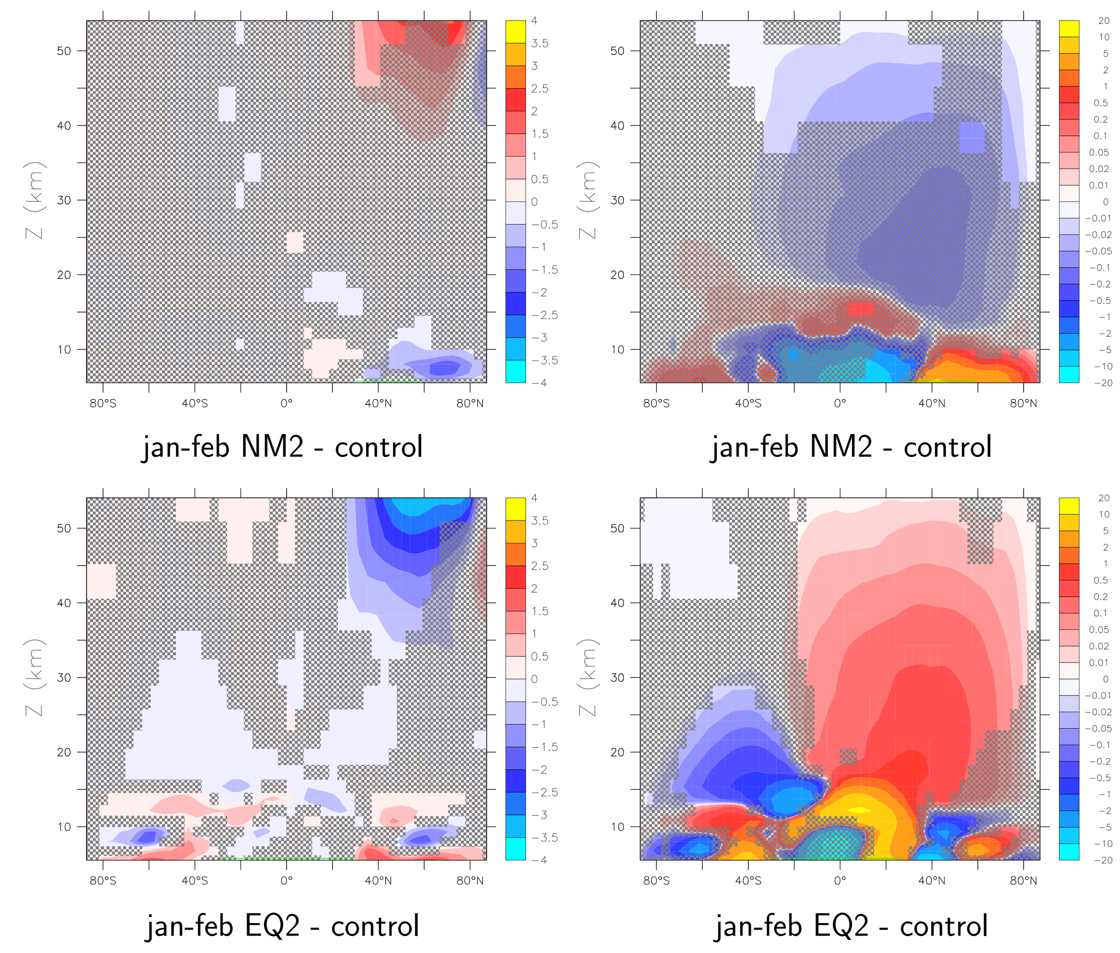

Fig 3: Left: Difference in EP flux divergence (10^{-5} m/s^2). **Right:** Difference in TEM streamfunction (μg/s).

- Top two panels, NM2: Divergence of the EP flux is seen in the vortex, implying decreased wave forcing of the mean flow, and the Brewer-Dobson circulation is correspondingly diminished. This is consistent with the stronger vortex seen on Figure 2.
- Bottom two panels, EQ2: There is significant convergence of the EP flux into the polar vortex, leading to wave-forcing of the mean flow there. The Brewer-Dobson circulation is consequently enhanced. This is consistent with the much weaker polar vortex seen on Figure 2.

Wave source regions (NM2, EQ2)

The control case isentropes overlaid on Figure 1 show that 30°N is the latitude of strongest baroclinicity in the NH, and the meridional paths of vertically propagating waves appear to be very sensitive to the surface temperature at this latitude.

EQ2 surface temperature forcing leads to increased wave forcing of the vortex. NM2 surface temperature forcing leads to decreased wave forcing and a stronger vortex.

We use the zonal-mean eddy heat flux $\overline{v'T'}$ to locate regions of vertical wave propagation. This quantity, when averaged over 40°N-80°N and over Jan-Feb at a height of 100 hPa is used as an indicator of the wave forcing in the vortex [4].

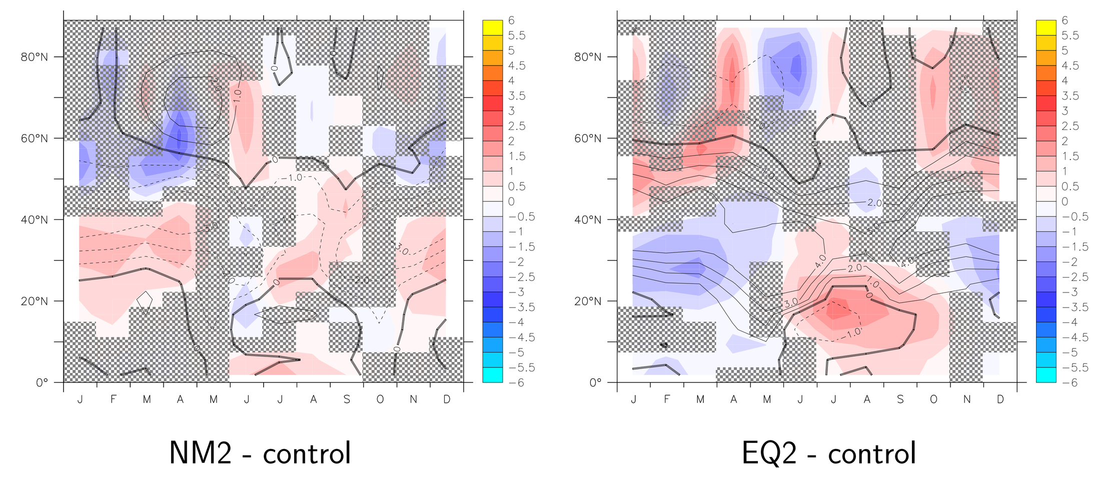

Fig 5: Zonal-mean eddy heat flux $\overline{v'T'}$ (K m/s) at 100 hPa. The contours give the difference in the zonal wind speed at the jet core.

- EQ2: There is decreased eddy heat flux south of 40°N for Jan-Feb, but increased values north of 40°N, suggesting increased wave activity in the polar vortex.
- NM2: The situation is reversed, with decreased eddy heat flux north of the 40°N mark, supporting the observed strengthening of the polar vortex. There is increased eddy heat flux south of 40°N, but the wave activity associated with it is directed elsewhere than into the vortex.
- In both cases, the areas of increased eddy heat flux coincide roughly with the areas of most diminished, or least enhanced, jet core windspeed. In the NM2 case, the tropospheric jet weakens and wave flux increases in the jet latitudes. In the EQ2 case, the jet becomes stronger, and wave flux decreases in the jet latitudes.

Conclusions

- Changing the location of the strongest tropospheric winds changes the direction in which waves will propagate into the stratosphere.
  Surface temperature forcing can strengthen or weaken the jet, depending on its location.
- The polar stratospheric temperature response (Figure 1) to NM2 and EQ2 forcing are equal in size and of opposite sign. Is the surface forcing effect in these bands additive?

References

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