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Supplement of

Circulation responses to surface heating and implications for polar amplification

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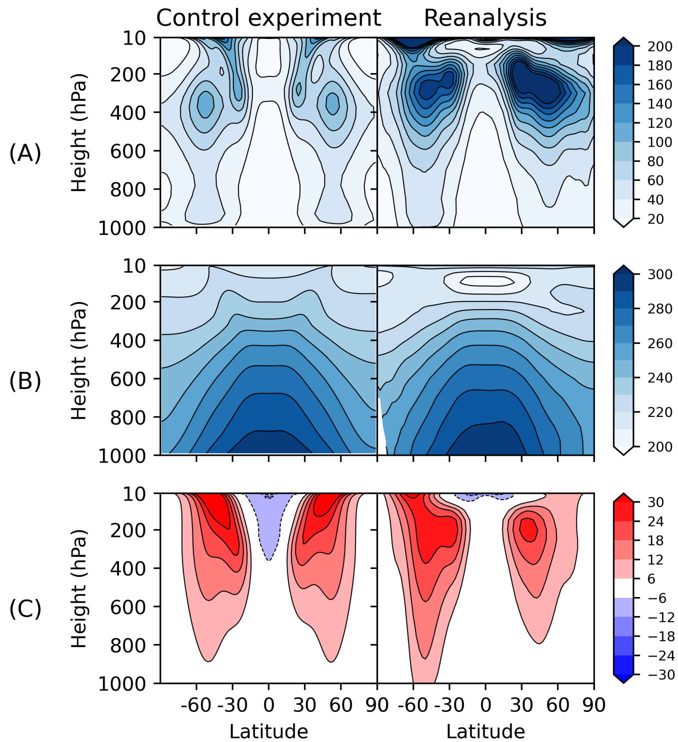

Figure S1. Annual, zonal mean of the **(A)** transient eddy kinetic energy (m^2s^{-2}) , **(B)** air temperature (K) and **(C)** zonal wind (ms^{-1}) for the control experiment (years 20-50; left column) and NCEP-DOE reanalysis 2 (years 1980-2010; right column) (Kanamitsu et al., 2002). The vertical axis shows the height; horizontal axis shows the latitude. The transient eddy kinetic energy is calculated as $\frac{1}{2}(u'^2 + v'^2)$, where u is the zonal wind, v is the meridional wind and prime (') is the daily deviation from climatology (equation 5.1 in James, 1995).

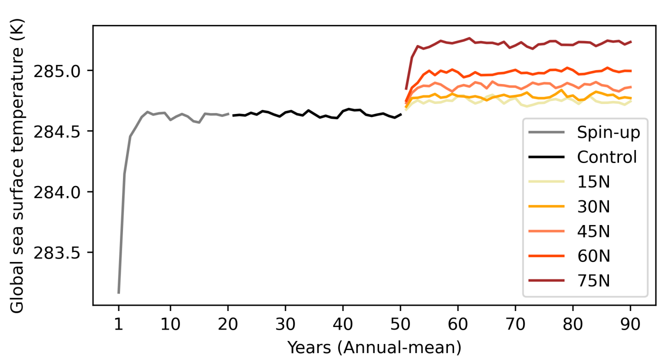

Figure S2. The temporal evolution of global, annual mean sea surface temperature (K) in the periods of spin-up (years 1 to 20), control (years 20 to 50) and the perturbation experiments (years 50 to 90). The perturbation experiments are branched off from the last day of year 50 from the control experiment. The first 10 years (years 50 to 60) from the perturbation experiments are discarded.

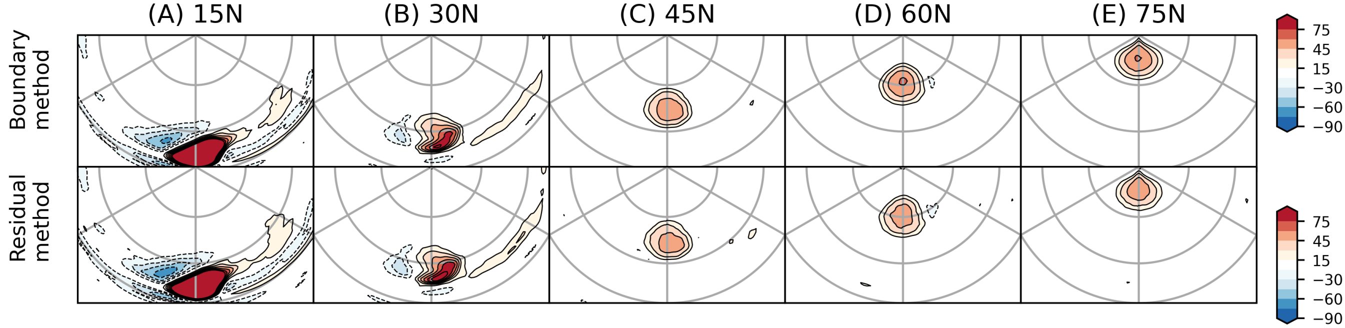

Figure S3. The column integrated diabatic heating response (Wm^{-2}) calculated by the boundary (top row) and residual methods (bottom row) in the **(A)** 15°N, **(B)** 30°N, **(C)** 45°N, **(D)** 60°N and **(E)** 75°N heating perturbation experiments. The column integrated diabatic heating using the boundary method is equal to $R_t - R_s + H_s + LP$, where R_t is the net radiative fluxes at the top of the atmosphere, R_s is the net radiative fluxes on the surface, H_s is the surface sensible heat flux, L (=2.26×10^6 Jkg^{-1}) is the latent heat of evaporation and P is the precipitation rate (see Equation 5 in Trenberth and Solomon, 1994). The column integrated diabatic heating using the residual method is equal to $\frac{C_p}{g}\int_{1000hPa}^{0hPa}\overline{Q}dp$, where C_p (=1004 $JK^{-1}kg^{-1}$) is the specific heat capacity of air at a constant pressure, g (=9.8ms^{-2}) is gravity of Earth, p is the pressure, and $\overline{Q}$ is the time-mean diabatic heating obtained as the residual from the thermodynamic equation (Equation 2 in the main text).

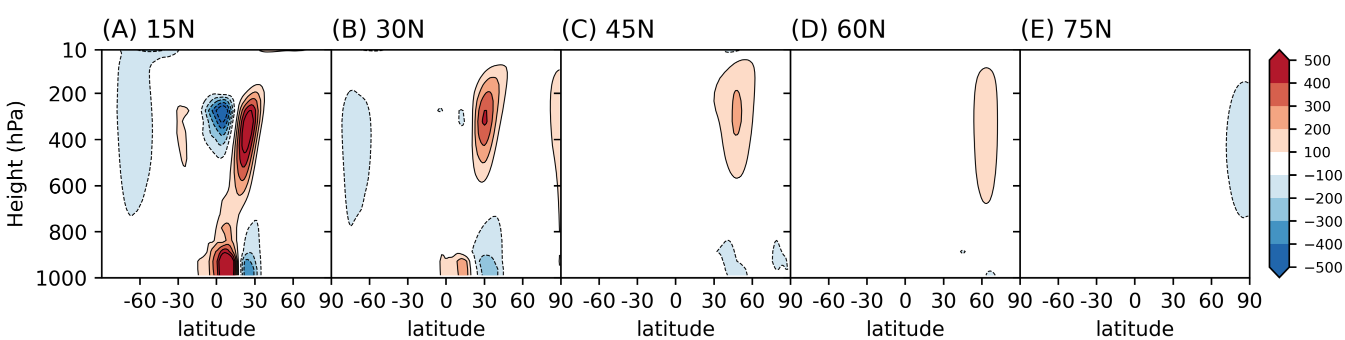

Figure S4. Meridional heat transport response ($\overline{vT}$, Kms^{-1}) in the **(A)** 15°N, **(B)** 30°N, **(C)** 45°N, **(D)** 60°N and **(E)** 75°N heating perturbation experiments. The latitude-height section is shown as the zonal average over the longitudes of the Q-flux perturbations.

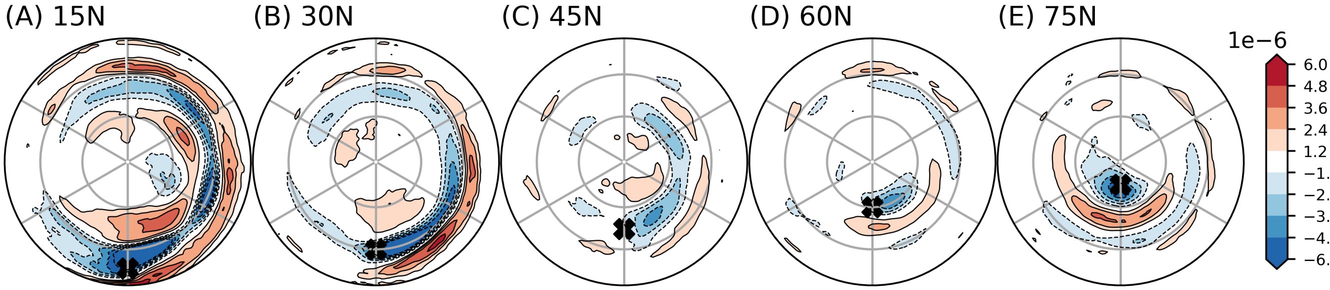

Figure S5. Upper level (232 hPa) vorticity response (s^{-1}) in the **(A)** 15°N, **(B)** 30°N, **(C)** 45°N, **(D)** 60°N and **(E)** 75°N heating perturbation experiments. The crosses mark the position of the surface heating perturbation. The latitude lines mark 30°N and 60°N. The longitude lines denote 60° intervals, marking 120°W, 60°W, 0°, 60°E, 120°E and 180°.

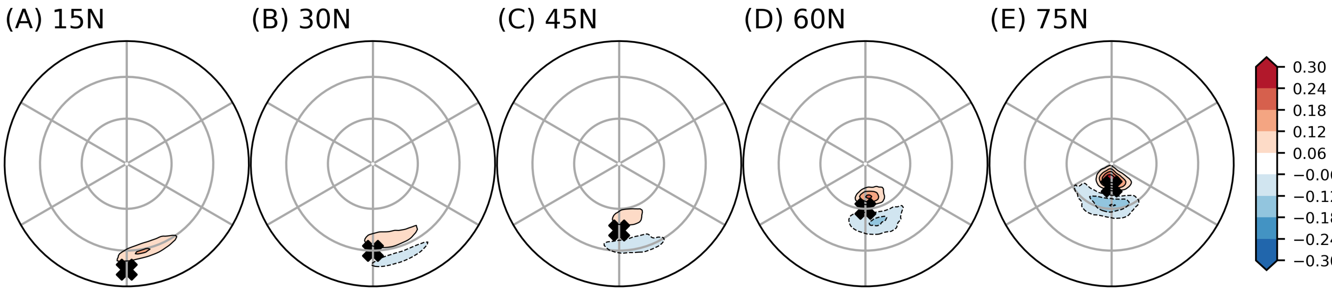

Figure S6. 780 hPa Eady growth rate maximum response (day^{-1}) in the **(A)** 15°N, **(B)** 30°N, **(C)** 45°N, **(D)** 60°N and **(E)** 75°N heating perturbation experiments. The crosses mark the position of surface heating perturbation. The Eady growth rate is calculated as $0.31|f||\frac{du}{dz}|N^{-1}$, where f is planetary vorticity, N is the Brunt-väisälä frequency and $\frac{du}{dz}$ is the vertical zonal wind shear (Hoskins and Valdes, 1990; Simmonds and Lim, 2009). The crosses mark the position of the surface heating perturbation. The latitude lines mark 30°N and 60°N. The longitude lines denote 60° intervals, marking 120°W, 60°W, 0°, 60°E, 120°E and 180°.

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