Overview of the dynamics of the Brewer-Dobson stratospheric circulation

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Based on Vallis' atmosphere-ocean fluid dynamics textbook

- Zonal stratospheric circulation (Vallis Fig 13.12, and p 568): shortwave (SW) absorption near summer pole leads to a reversed temperature gradient in summer hemisphere: e.g., $T_y < 0$ in southern hemisphere during Jan. This leads to $u_z \propto T_y/f_0 > 0$ in southern hemisphere July, and using $u = 0$ at top of stratosphere we get $u < 0$ (easterlies) during summer (Jan) in the southern hemisphere stratosphere. Similarly, $u < 0$ (easterlies) during summer (July) in northern hemisphere.

- Winter hemisphere (northern Jan, southern July) has no SW at pole, temperature gradient is not reversed and winds are westerlies there.

- Rossby waves propagating vertically from the troposphere cannot propagate into easterlies, therefore can only reach stratosphere in the winter hemisphere.

- Brewer-Dobson stratospheric circulation: zonally averaged momentum balance is $-f_0\overline{v}^* = \overline{v'q'}$ (Vallis, eqn 13.88; $q' = \zeta' + f\partial_z(b'/N^2)$, see chapter 7.2). Assuming the potential vorticity flux is down gradient (equatorward, because the gradient is dominated by β), the rhs is negative, so that the mean flow $\overline{v}^* > 0$ is poleward.

- B-D circulation warms the pole and cools the equator in the stratosphere (Vallis eqn 13.89): $N^2\overline{w}^* = \frac{\theta_E - \theta}{\tau}$, together with positive w in tropics and negative in polar areas forced by poleward B-D meridional flow. This leads to $\theta < \theta_E$ (cooling!) at equator (where $\overline{w}^* > 0$) and $\theta > \theta_E$ (warming!) at pole (where $\overline{w}^* < 0$).