Dynamics in the stratosphere

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Contents

1 Introduction

In this set of notes I'll cover the mean state and variability in the stratosphere. The radiative properties of the stratosphere aren't covered here (see notes on radiative transfer for that)

2 The stratosphere mean state and the Brewer-Dobson circulation

To understand the Quasi-biannual oscillation (QBO), it's helpful first to have an idea of the zonally averaged mean meridional circulation in the stratosphere and of the sources of variability around that mean. We'll begin by inspecting the zonally averaged fields of temperature and zonal wind as a function of latitude and height (see Fig. 1).

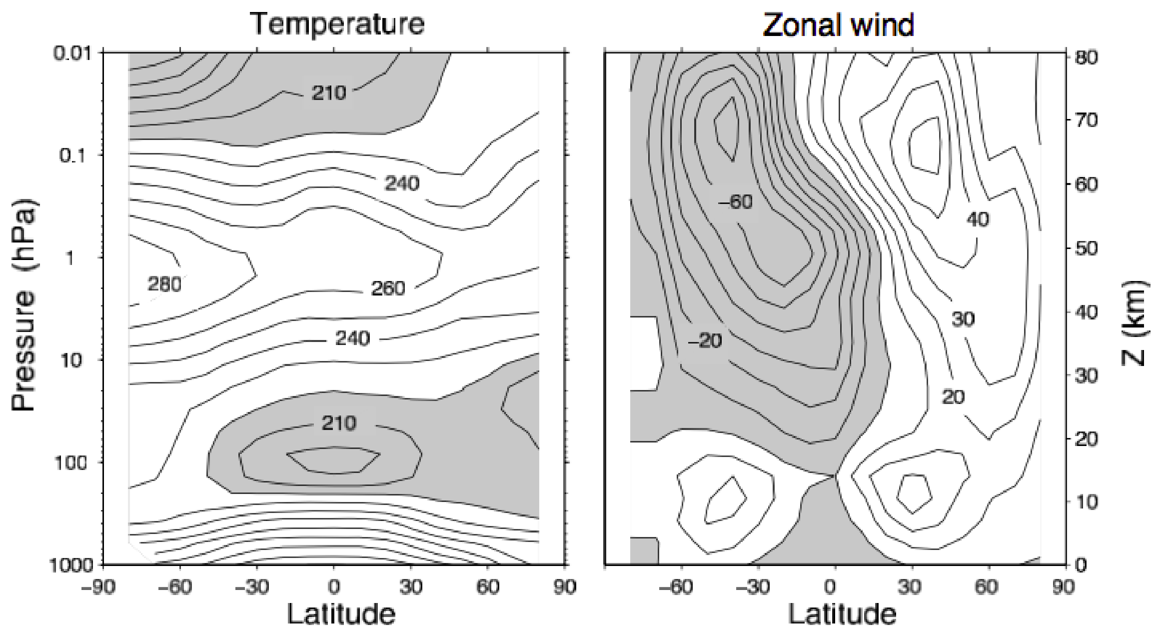

Fig. 13.12 The zonally averaged temperature and zonal wind in January. The temperature contour interval is 10 K, and values less than 220 K are shaded. Zonal wind contours are 10 m s^{-1} and negative (westward) values are shaded.[21]

Figure 1: Figure credit: Chapter 13 of (Vallis, 2006)

To understand how the temperature relates to the zonal wind, we need to remember the thermal wind equation:

$$\frac{\partial u_g}{\partial z} = -\frac{R}{fH}\frac{\partial T}{\partial y}$$

in words: the vertical gradient of the zonal wind is inversely related to the meridional gradient in temperature. So in the summer hemisphere (the southern hemisphere) stratosphere at ~1 hPa, where there is a strong positive meridional gradient of temperature from the equator to the pole in January, the zonal wind decreases (becomes westward) with height. In the winter hemisphere (the NH) the temperature gradient at $\sim 1hPa$ is negative, and so the zonal wind increases in strength with height from the jet stream into the stratosphere.

Figure 1 may therefore imply that thermal wind alone explains the zonal winds. But if we compare observations to a simple temperature profile from radiative equilibrium above 20 km (40 hPa) we can see that the meridional gradient in temperature is actually less intense than would be expected, which means that large-scale dynamics must be working to smooth out the meridional temperature gradient set up by radiative equilibrium (see Fig. 2).

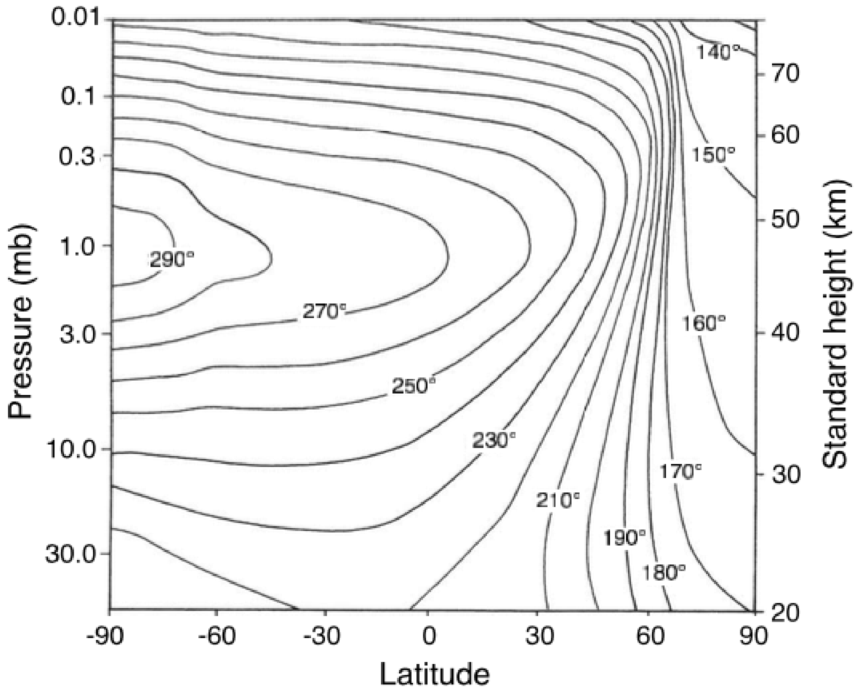

Fig. 13.11 The zonally averaged radiative-equilibrium temperature in January. The downward solar radiation at the top of the atmosphere is given, and the upward radiative flux into the stratosphere is based on observed properties, including temperature, of the troposphere.[20]

Figure 2: Figure credit: Chapter 13 of (Vallis, 2006)

So let's step back and summarize: The stratosphere is stably stratified in the vertical (in part due to O_3 absorption between 20 and 50 km). Meridionally, the summer hemisphere pole is warmer than the equator (again, absorption by O_3) such that, by thermal wind, the zonal wind decreases with height. In the winter hemisphere the equator-to-pole temperature gradient is flipped and so zonal winds increase in strength from the jet stream up to the stratosphere. The temperature in the stratosphere resembles a damped version of what would be predicted by radiative equilibrium, meaning that large-scale dynamics are important in damping meridional temperature gradients.

The large-scale circulation in the stratosphere is referred to as the Brewer-Dobson circulation. It refers to a single equator-to-pole cell in each hemisphere, which is stronger in the winter hemisphere than the summer hemisphere. The Brewer-Dobson circulation represents contributions from both the Eulerian mean and eddy-driven components of the circulation. Figure 3 depicts the Brewer-Dobson circulation in schematic form: In the tropics air is drawn up from the troposphere, it then turns polewards before sinking at high latitudes.

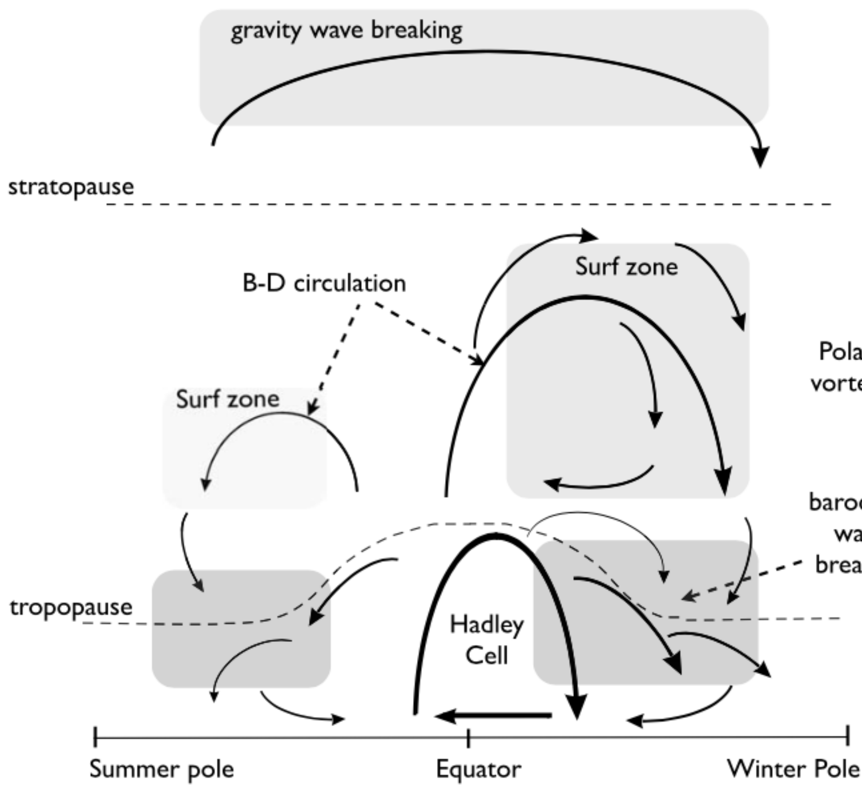

Fig. 13.14 A schema of the residual mean meridional circulation of the atmosphere. The solid arrows indicate the residual circulation (B-D for Brewer–Dobson) and the shaded areas the main regions of wave breaking (i.e., enstrophy dissipation) associated with the circulation. In the surf zone the breaking is mainly that of planetary Rossby waves, and in the troposphere and lower stratosphere the breaking is that of baroclinic eddies. The surf zone and residual flow are much weaker in the summer hemisphere. Only in the Hadley Cell is the residual circulation comprised mainly of the Eulerian mean; elsewhere the eddy component dominates.

Figure 3: Figure credit: Chapter 13 of (Vallis, 2006)

The main source of wave activity in the stratosphere is upward propagation from the troposphere, where zonal asymmetries in heating, topography or baroclinic instability may create wave activity. It's important to remember here that for planetary-scale Rossby waves to propagate up into the stratosphere, the zonal mean flow needs to be westerly (eastward). This is shown in Figure 4.

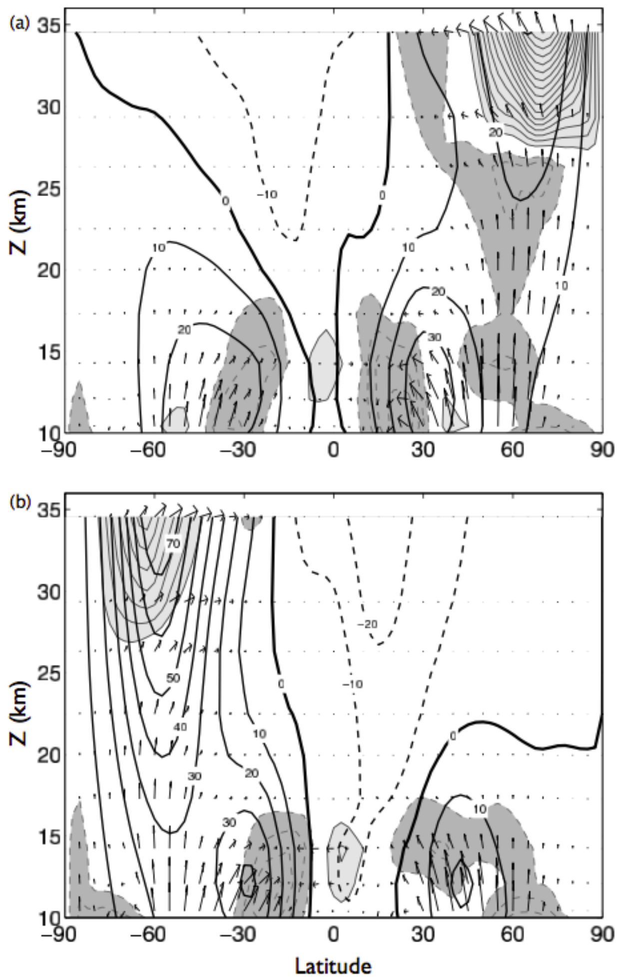

Fig. 13.15 The EP flux vectors (arrows), the EP flux divergence (shaded and light contours) and the zonally averaged zonal wind (heavy contours) for (a) northern hemisphere winter; (b) southern hemisphere winter. Note the almost zero EP values in the summer hemispheres, and strong convergence at high latitudes in the winter hemispheres, leading to poleward residual flow and/or zonal flow acceleration. The EP divergence is shaded for values greater than $+1\ \mathrm{m\,s^{-1}}$/day, light solid contours) and for values less than $-1\ \mathrm{m\,s^{-1}}$/day (light dashed contours). The vertical coordinate is log pressure, extending between about 260 and 10 mb.

Figure 4: Figure credit: Chapter 13 of (Vallis, 2006)

Another way to consider why flow in the stratosphere is equator-to-pole is to consider the circulation as a result of breaking Rossby waves, which are ‘pumped’ up to the stratosphere in the tropics. The Rossby waves break in the equatorial stratosphere, depositing westward momentum (wave drag), which is largely balanced by the Coriolis force that will act to the right (poleward; see Fig. 3).

3 The Quasi-Biannual Oscillation (QBO)

Having this mean circulation in mind, we can now consider variability in the stratosphere (in particular variability in the zonal flow at the equator. The dominant internal oscillation in the stratosphere is the Quasi-biannual oscillation (QBO), which is an equatorially confined oscillation that describes the reversal of zonal winds every ∼24-29 months. These zonal reversals propagate downward through the stratosphere, as depicted in Figure 5. So what creates this reversal, what determines its period, and why does it propagate downward? In general to answer these questions we need to acknowledge that vertically-propagating waves have **both** negative and positive phase speeds, and that when they break in the stratosphere they affect the mean-flow.

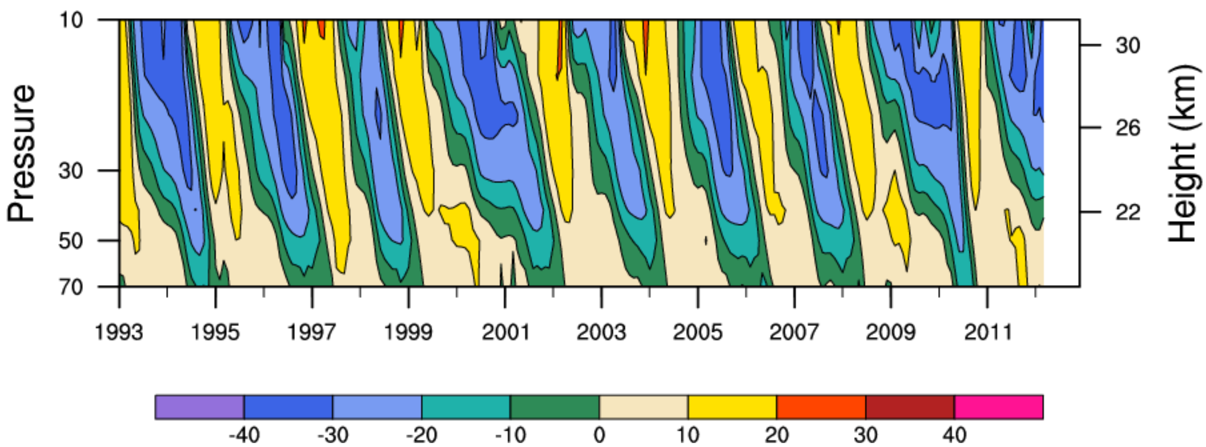

Figure 5: Figure credit:

Key to understanding how the QBO operates is wave-mean flow interaction, and the notion of a ‘critical level’, where the wave phase speed is equal to the mean zonal wind. In the presence of shear a wave will be attenuated more quickly near it’s critical level. This means that waves will preferentially break as they approach the critical level. Provided that waves with a negative phase speed will propagate unimpeded through westerly flow while waves with a positive phase speed will propagate through easterly flow, we can investigate how the zonal mean flow (as a function of height) changes as a result of wave breaking.

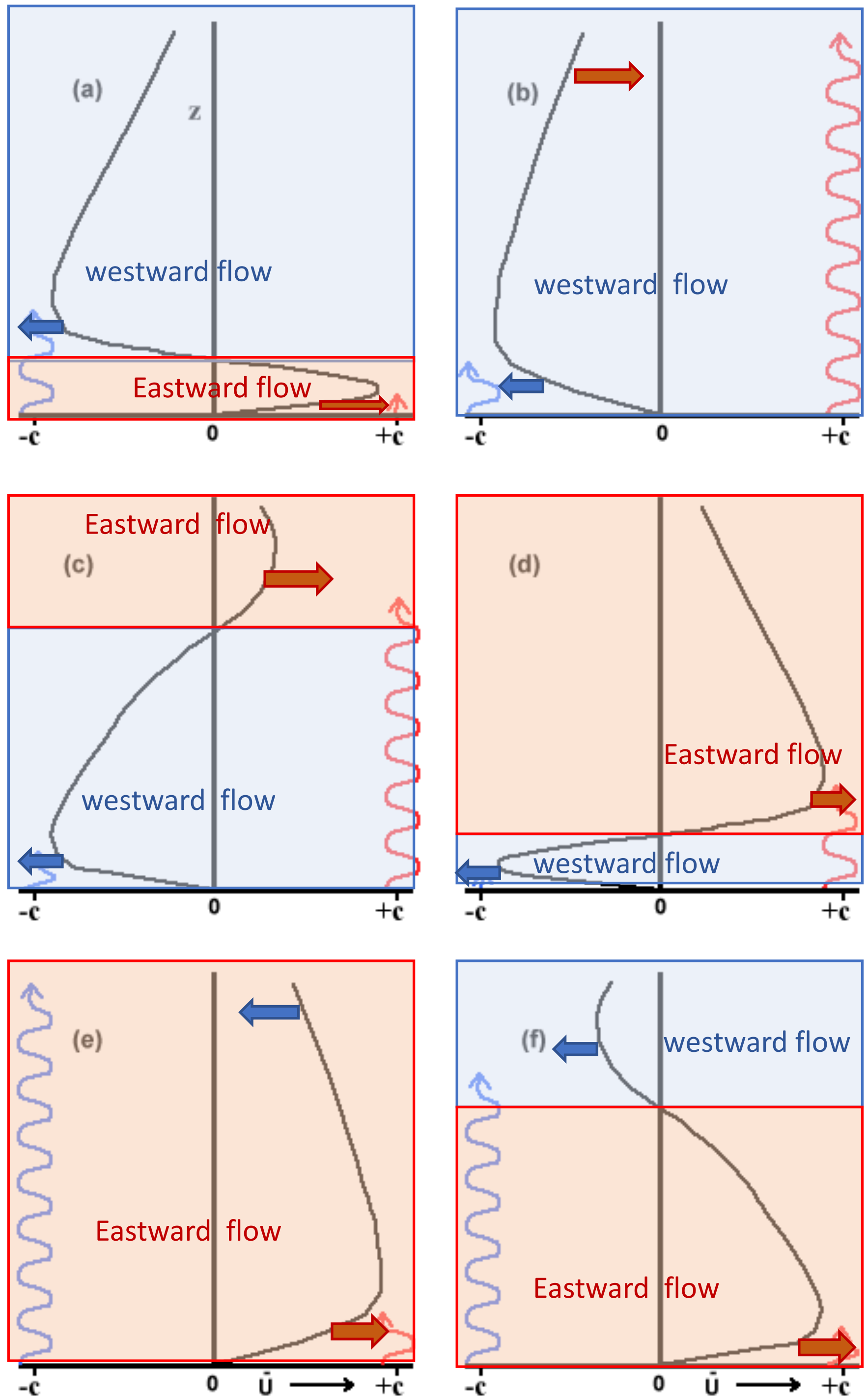

Figure 6: The solid line represents the mean flow as a function of height, while the wavy blue or red lines indicate vertical propagation of waves. The inflection points in the solid line represent the critical zones, where waves preferentially break. Waves with a negative (positive) phase speed deposit westward (eastward) momentum into the mean flow.

Figure 7 demonstrates how breaking waves act to reverse the zonal flow in the stratosphere. When a wave approaches the critical level it breaks and deposits momentum into the mean flow, accelerating the mean flow towards its wave speed and displacing the critical level down towards the wave-breaking. Importantly, this is happening in two opposite directions at two separate heights, which creates the zonal shear. In panel (a) of Fig. 7 the easterly flow zone (westerly jet), in which waves with a positive phase may propagate, is descending until it reaches the wave source (beyond which it can't descend) where it is dissipated due to the extreme shear. The zonal flow then (in (b)) is entirely easterlies, where positive phase waves propagate and break high in the stratosphere, exhibiting drag near the top of the profile. Meanwhile the negative phase waves continue to pull the critical point downward in the lower stratosphere (panel b). This process continues in panels c-f and resumes in panel a having completed a full cycle. Importantly, note that the period of the zonal-mean wind (and therefore the QBO) depends on the amplitude (not period) of the breaking waves.

So, turning to our final question, why is the QBO confined to the equator? As we briefly mentioned in our discussion of the mean flow, waves propagate into the stratosphere from the mid-latitude troposphere as well as the tropics. However, off the equator the wave-breaking can be balanced by the Coriolis force. Near the equator wave-breaking leads to a zonal acceleration. This is why the QBO is confined to the equatore (where the Coriolis force is small).

Consider a final figure summarizing propagation of planetary waves in the mid-latitudes and propagation of waves in the troposphere, which collectively drive zonal flow in the stratosphere.

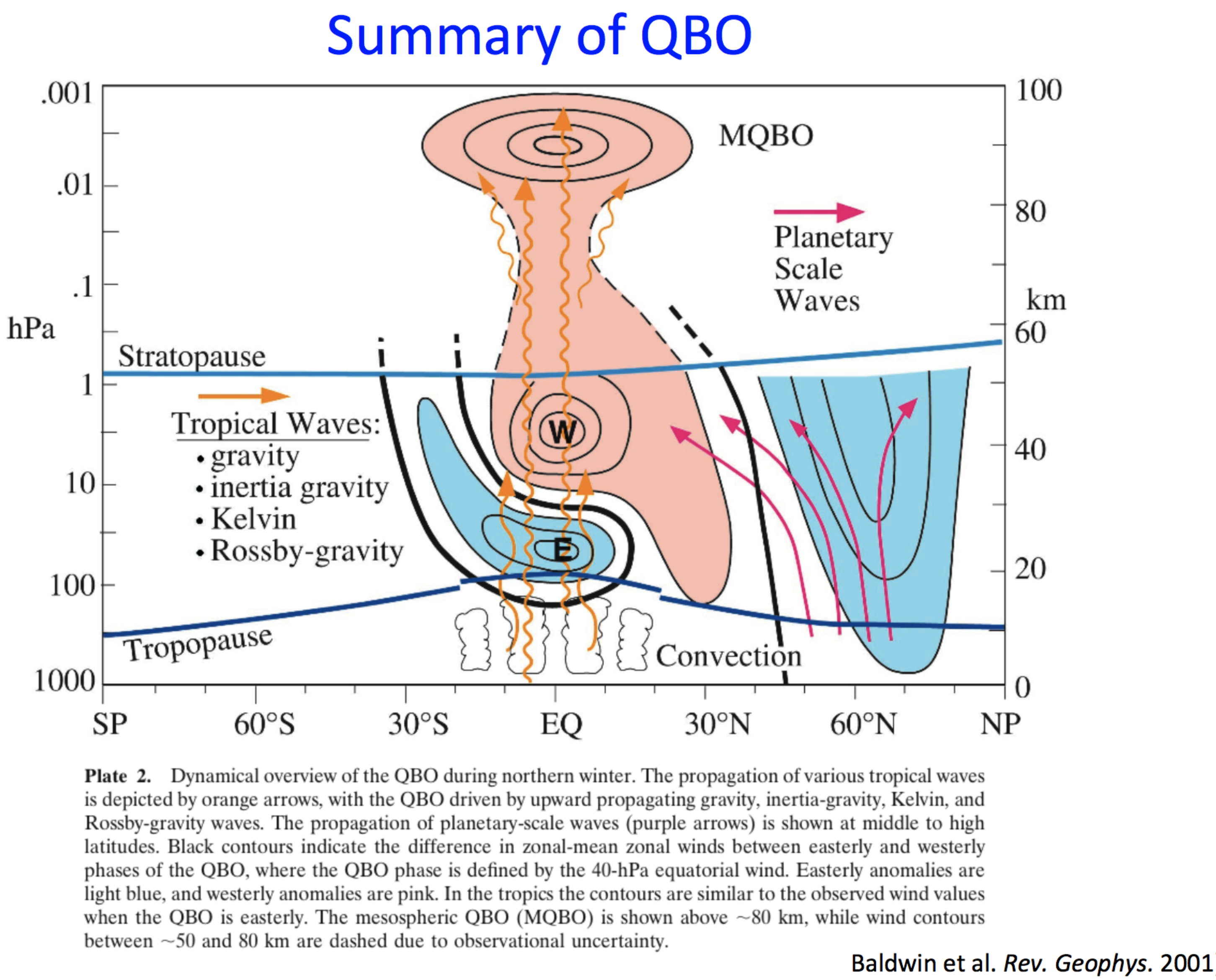

Figure 7: Figure Credit: Lecture notes of Ron Miller, NASA GISS

References

Geoffrey K Vallis. *Atmospheric and oceanic fluid dynamics: fundamentals and large-scale circulation.* Cambridge University Press, 2006.