Tropical Tropopause Layer (TTL)

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I- The tropopause

There are many studies trying to understand and define exactly what the tropopause is, and which method is the best for particular regions of the globe (e.g. Reed, 1955; Danielsen, 1959; Shapiro, 1980; Danielsen et al., 1987; Hoerling et al, 1991; Hoinka, 1998). In general, the tropopause is the transition layer between the troposphere and the stratosphere, where an abrupt change in temperature lapse rate usually occurs. Following the World Meteorological Organization (WMO), it is defined as the lowest level at which the lapse rate decreases to 2 K/km or less, that is, the average lapse rate between this specific level and all higher levels within 2 km does not exceed 2 K/km.

Tropopause height shows large variations with latitude, season, and even from day-to-day. Latitudinal variation of the tropopause from the poles to the equator is schematically illustrated in Fig. 1, and Table 1 shows some differences between the tropical and extratropical tropopauses.

Although the lapse rate is a method to identify the tropopause, it is useful to use the concept of a 'dynamical tropopause' in the extratropics through potential vorticity surfaces. However, within the tropics the thermal definition based on the stability is the only applicable criterion. Hoinka (1998) used both methods to get a global tropopause, as shown in Fig. 2.

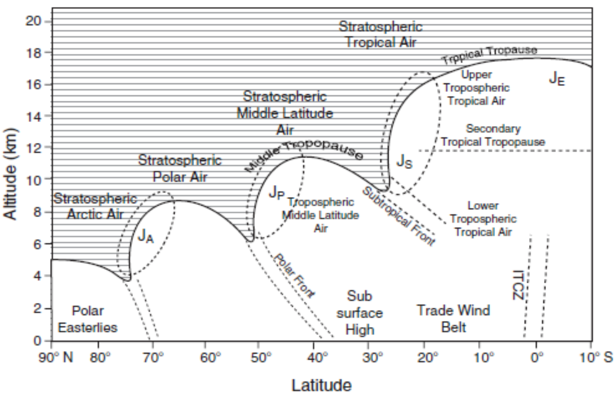

Table 1 Characteristics of tropical and extratropical tropopauses

Fig. 1 Scheme of tropopause according to different latitudes. Adapted from Shapiro et al. 1987.

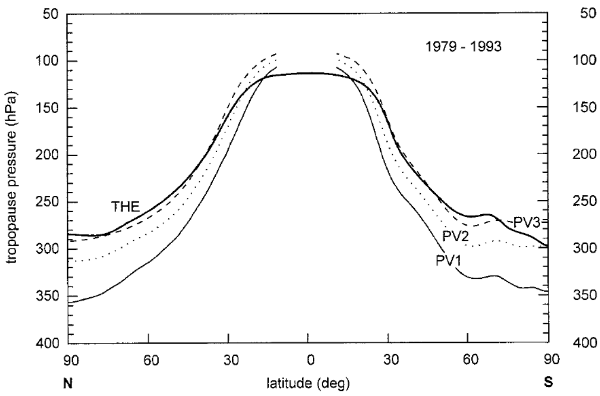

Fig 2. Annual mean meridional profile of the zonally averaged tropopauses (1979-93; 1200UTC): thermal tropopause (THE); dynamical tropopause with 1.6 (PV1), 2.5 (PV2), and 3.5 (PV3) PVU.

Tropical (TT)	Extratropical (ET)
Colder (-80°C)	Warmer (-60°C)
In the tropics, between the two subtropical jet streams	Between subtropical and polar jet streams
~ 18 km (80-100 hPa)	~ 12 km (200-300 hPa)
Sharply defined, highest and coldest	Higher in summer and lower in winter
Radiative-convective balance	Baroclinic wave dynamics
Upward circulation	Downward circulation

II- What is the Tropical Tropopause Layer (TTL)?

The tropopause in the tropics is defined by the layer between the level of maximum convective outflow (10–12 km) (closely corresponding to a minimum in ozone) and the Cold Point Temperature (CPT) (16–17 km) (Gettelman and Forster, 2002), as can be seen in Fig. 3.

More precisely, at the CPT, the altitude of the tropical tropopause layer (TTL) upper boundary is nearly uniform throughout the tropics, but shows a pronounced annual cycle (Randel and Jensen, 2013). This annual cycle is observed in Fig. 4. In addition, the TTL lower boundary height has regional variations (Gettelman et al., 2002). Furthermore, the interannual variations of the TTL result from changes in the large scale organization of convective activity, such as the annual cycle in the strength of upwelling in the Brewer-Dobson circulation, the descending zonal wind regimes in the quasi-biennial oscillation (QBO), and the interannual El Niño-Southern Oscillation (ENSO) (Gettelman et al., 2002).

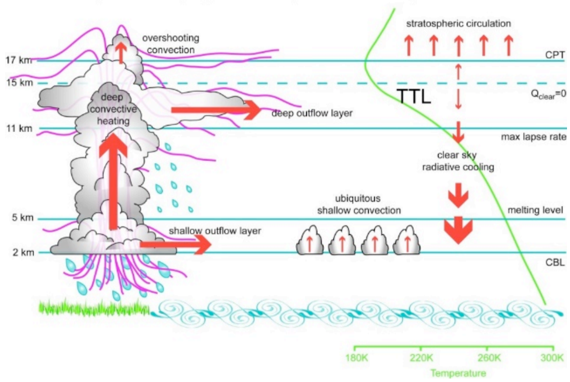

Fig. 3 Tropical Tropopause Layer and Deep Convection (Birner, 2008)

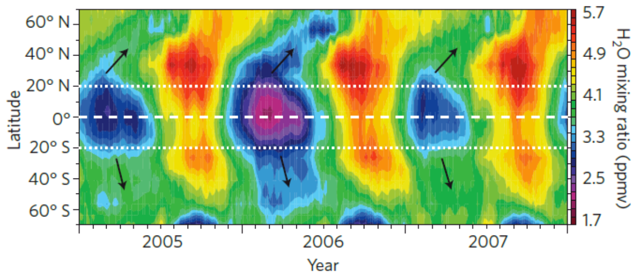

Fig. 4 Satellite Observation of water vapor in the lower stratosphere demonstrates the transport of air from the TTL to higher latitudes (Randel et al., 2013).

This transition zone allows troposphere to stratosphere transport (TST). Convective activity also has an important impact on the TST of water vapour. More precisely, the TST is very important in the tropics (Brewer, 1949; Fueglistaler et al., 2009). Once in the stratosphere, WV is transported into the winter hemisphere from the tropics upward and towards the extra-tropics (Potter and Holton, 1994) (illustrated in Fig. 4). This phenomenon is called the Tape Recorder effect (Mote et al., 1996) and is evidence for the Brewer-Dobson circulation. Furthermore, global atmospheric water vapour is controlled by the TTL temperature: the TTL water vapour decreases with the TTL temperature. Thus, TTL temperature plays a significant role in controlling the global stratospheric water vapour, which contributes to ozone destruction in the polar regions. However, stratospheric dehydration and hydration are different processes, which happen on different time scales (Sherwood and Dessler, 2000), and are still debated.

III - Its significance in transporting water vapor to the stratosphere

The regions of coldest temperatures in the TTL play a crucial role in setting the boundary condition to water vapour transport into the stratosphere. Previous studies on CPT structure and variability reveal that there is longitudinal asymmetry in CPT temperatures (Kim and Son, 2012). The coldest temperatures are observed over the regions where frequent deep convective events occur. Nevertheless, it is important to note that horizontal transport time scales in the TTL are approximately four orders higher than the vertical transport time scales (Holton and Gettelman, 2001). Hence, the dehydration of air in the TTL depends on both the slow vertical ascent in the tropics as well as the horizontal transport through the coldest regions (e.g. equatorial western Pacific).

In the tropics, the air temperature at the CPT is extremely cold. "*Air transiting the TTL is substantially dehydrated on passing the cold tropical tropopause, leading to the extreme dryness (just a few ppm of water vapour) of the global stratosphere*" (Randel and Jensen, 2013; Brewer, 1949). Thus at the CPT level (~100 hPa) we expect water vapour concentration which is saturated for local CPT temperatures. For the present study, we used CPT estimated from Radio Occultation data, and water vapour measured from the Aura-Microwave Limb Sounder (MLS). Here we have considered the monthly mean data of CPT and water vapour at a spatial resolution of 2.5°x 2.5°.

The saturation of water vapour concentration calculated for CPT temperature over the western Pacific (lat: 5°S-5°N, lon: 130°E-150°E) is represented as the black line in Fig. 5. The blue and green lines represent observed water vapour concentration over the western Pacific at 100 hPa and 82 hPa pressure levels, respectively. The inter-annual variation of water vapour concentration is well correlated (~0.9) with water vapour saturation value at the CPT. The tape recorder effect is reflected in Fig. 5 as the lag between the blue and green lines.

It is expected that the regions with higher correlation between observed water vapour at 100 hPa and saturation water vapour concentration at CPT level probably contribute more to lower stratospheric water vapour. Thus, similar analysis that has been carried out for the western Pacific is extended to all longitudes (averaged for 5°S-5°N) and is depicted in Fig. 6. Higher correlations are clearly observed over all longitudes in the equatorial region except in the eastern Pacific and west equatorial Indian Ocean. These two regions where we observe relatively weak correlations are also the locations of sinking motion (in the upper troposphere) associated with zonal Walker Circulation. Over the descending regions of Walker Circulation (eastern Pacific and west equatorial Indian Ocean), the interannual variability in water vapour concentrations at CPT altitudes (~100 hPa) are not well related to its interannual variability observed in the CPT temperature over these regions.

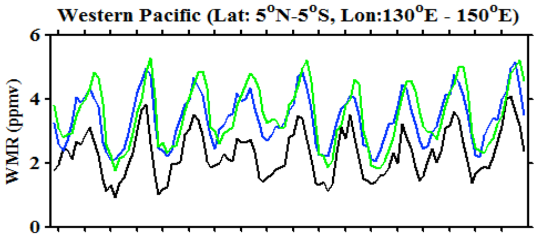

Fig. 5 MLS measured water vapor mixing ratio at 100hPa (blue line) and 82 hPa (green line). The water vapor saturation mixing ratios estimated for CPT temperatures are represented by black line.

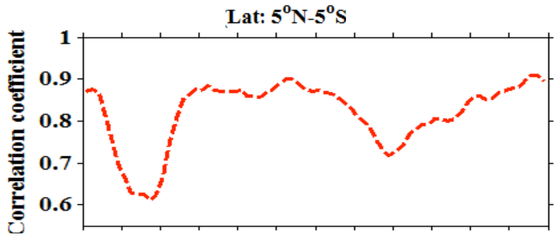

Fig. 6 Correlation coefficients between observed water vapor at 100 hPa and water vapor saturation value at each CPT temperatures (Suneeth and Das, 2017).

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