Variation in the tropopause transition layer over China through analyzing high vertical resolution radiosonde data

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ABSTRACT
In this work, the meridional variation of the thermal tropopause layer is investigated using high vertical resolution radiosonde measurements over China. A tropopause-based mean method is used to characterize the depth of the tropopause inversion layer (TIL), and a curve-fitting method to profiles of Brunt–Vaisala frequency is applied to identify the transition sharpness of the static stability in the tropopause transition layer (TTL). The eight radiosonde stations are grouped into four latitudinal bands at intervals of 7°. Analyses show that the tropopause at higher latitudes is much more likely to be sharp and have a strong TIL, while that at lower latitudes is dominated by a thick TTL.

1. Introduction
The tropopause, as the top of the troposphere or the boundary between the troposphere and stratosphere, was coined very naturally after the discovery of the stratosphere at the end of the 1890s (Hoinka 1997). Since its birth, the definition of the tropopause has evolved with the continuous enhancement and capability of atmospheric observations along with growing interest from scientists in different aspects of atmospheric processes (Bian 2009; Fueglistaler et al. 2009; Gettelman et al. 2011). In the early model of atmospheric thermal structure, a formal definition of the thermal tropopause was confirmed based on a temperature lapse rate of 2 K km⁻¹ (WMO 1957). Later, a dynamical concept, potential vorticity, was used to define the tropopause (Danielsen et al. 1987; Hoskins 1991). Although both thermal and dynamical definitions of the tropopause show general agreement in terms of its large-scale distribution, there are certain differences in between them in terms of tropopause height on smaller scales, particularly in the vicinity of jets (Hoerling, Schaack, and Lenzen 1991). The differences arise from the underlying distinction in physical meaning: the thermal definition is concerned with the transition in the vertical thermal structure, and the change in static stability is recognized as the main indicator; whereas, the dynamical tropopause involves finding a quasi-mass surface to define the chemical transition from stratosphere to troposphere (Shapiro 1980). Thereafter, a few new definitions based on chemical species, such as ozone and water vapor, have been issued (Bethan, Vaughan, and Reid 1996; Pan et al. 2004; Zahn, Brenninkmeijer, and van Velthoven 2004)

Originally, the tropopause was considered to be a transport barrier between the troposphere and stratosphere. During the past few decades, however, scientists have recognized that this interface is not always clear, and the air-mass at the boundary between the troposphere and stratosphere sometimes possesses characteristics of both layers. Birner, Dornbrack, and Schumann (2002)
presented climatological temperature profiles relative to the tropopause based on radiosonde data with high vertical resolution over two midlatitude sites in southern Germany. A tropopause inversion layer (TIL) is frequently observed in temperature profiles immediately above the tropopause (Birner 2006; Birner, Dornbrack, and Schumann 2002). When a TIL is present, the tropopause is usually very sharp, with an abrupt transition between the tropospheric and stratospheric lapse-rate. Wirth (2003) proposed that large-scale dynamic processes can explain the presence of such a strong TIL in extratropical regions. However, the radiative transfer calculation indicates that strong vertical gradients in both ozone and water vapor near the tropopause contribute to the formation of the TIL (Randel, Wu, and Forster 2007). So, the formation mechanism of the TIL is still a topic under investigation (Gettelman et al. 2011).

The structure of the tropopause has been investigated in Europe and America (Birner 2006). However, no such studies have been conducted over Asia. Accordingly, in this study, we analyze the structure of the tropopause based on temperature profiles from high vertical resolution radiosonde data over China. Following this introduction, Section 2 describes the data-set and analysis methods. The tropopause-based mean thermal structure is described in Sections 3.1, and 3.2 presents an alternative method to characterize the transition in atmospheric stability near the tropopause. The results are summarized in the final section.

2. Data and method

2.1. Radiosonde data

In this work, high vertical resolution data were gathered from eight radiosonde stations over mainland China, covering 27° of latitude from 22°N to 49°N. The latitudes and longitudes for these eight stations are listed in Table 1. The stations are located in the eastern part of China, characterized by broad plains and hilly areas, which is collectively referred to by Chinese geologists as the Third Step of China's terrain. Therefore, the results derived from these measurements will demonstrate the features of the tropopause over the eastern part of China. The data time period spans 1 January 2009 through 31 October 2011. Radiosondes are typically available twice daily at 0000 UTC and 1200 UTC, and sometimes at 0600 UTC and/or 1800 UTC. The vertical resolution for each profile is ~10 m. In order to minimize the impact of temperature observational error on the lapse rate calculation, the data were sub-sampled to a lower vertical resolution of 50 m, which is found to be most efficient. Only data below 20 km in altitude are used, which is a suitable top for tropopause investigation (WMO 1957).

2.2. Thermal tropopause definition

According to the guidance of WMO (1957), the thermal tropopause is defined by the temperature lapse rate. The level of the tropopause is defined as the lowest altitude at which the temperature lapse rate decreases to below 2°C km⁻¹, provided that the average lapse rate from this level to any point within the next 2 km above does not exceed 2°C km⁻¹. The algorithm is applied only to altitudes above the 500 hPa level to avoid boundary layer inversions.

2.3. Tropopause-based averaging

Conventionally, the climatological vertical structure of temperature or static stability is calculated by sea-level-based averaging, which usually smooths some specific features around the tropopause and does not show the abrupt change of static stability there clearly. In order to highlight the finer structure, a tropopause-based method was introduced by Birner, Dornbrack, and Schumann (2002) to derive the seasonal mean thermal structure. In this method, the tropopause itself is considered to be a common reference level for each temperature profile, and all profiles are averaged with respect to the time-dependent tropopause level. Therefore, $H = z - z_{TP} + \bar{z}_{TP} (z_{TP}$ denotes tropopause altitude) is used as the vertical coordinate for the tropopause-based mean, where $z$ is the altitude with respect to sea level and $\bar{z}_{TP}$ is the seasonal mean tropopause altitude. Using this method, the TIL can be clearly highlighted; otherwise, this inversion structure would be blurred by the sea-level-based mean (Birner 2006; Birner, Dornbrack, and Schumann 2002). The thickness of the TIL in the tropopause-based mean profile is defined as the distance from the lower boundary (i.e. the tropopause) to the upper boundary, where the squared buoyancy frequency $N^2$ decreases to its typical stratospheric value.

2.4. Tropopause transition layer (TTL)

It is known that the average temperature lapse rate in the troposphere is about 6.5 K km⁻¹, whereas the temperature

Table 1. Radiosonde station information, and fraction of transition depth ($\lambda$) > 2 km.

| Station (abbreviation) | Latitude (°N)/longitude (°E) | Broad transition fraction (%) |
|-----------------------|------------------------------|-------------------------------|
| Nenjiang (NJ)         | 49.10/125.14                | 9.4                           |
| Xilinhaote (XLHT)     | 43.57/116.07                | 16.8                          |
| Beijing (BJ)          | 39.48/116.28                | 31.1                          |
| Zhengzhou (ZZ)        | 34.43/113.39                | 36.0                          |
| Nanyang (NY)          | 33.02/112.35                | 39.6                          |
| Yichang (YC)          | 30.42/111.18                | 46.9                          |
| Huaihua (HH)          | 27.33/110.00                | 42.6                          |
| Nanning (NN)          | 22.38/108.13                | 43.8                          |
increases with altitude in the stratosphere, which means that the lapse rate there is negative. So, there is a transition from weak static stability in the troposphere to strong stability in the stratosphere, and the transition sharpness reflects different dynamical and radiative processes in this altitudinal range (Randel, Wu, and Forster 2007; Wirth 2003). However, the thermal tropopause definition, based on a threshold of temperature lapse rate (2°C km⁻¹), does not show the sharpness of the tropopause or the depth of the TTL. To characterize the sharpness of the tropopause, a smooth step-like function is fitted to the observed profile of static stability N by using a nonlinear curve fitting method (Homeyer, Bowman, and Pan 2010; Markwardt 2009). The fitting function has the form:

\[
N(z) = N_{\text{trop}} + \frac{N_{\text{strat}} - N_{\text{trop}}}{2} \left( 1 + \tanh \left( \frac{2(z - z_0)}{\lambda} \right) \right),
\]

where \(N_{\text{trop}}\) and \(N_{\text{strat}}\) are the typical values of N in the troposphere and the stratosphere, respectively; tanh represents the hyperbolic tangent function. In this case, the tropopause level is taken to be the transition midpoint of \(z_0\), and the depth of the TTL from the troposphere to the stratosphere is denoted as \(\lambda\). The four parameters \(N_{\text{trop}}\), \(N_{\text{strat}}\), \(z_0\), and \(\lambda\) are calculated for each profile by using a least-squares fitting method.

Two typical cases are shown in Figure 1 to demonstrate the difference between the lapse-rate based tropopause and the TTL. For a case with relatively sharp transition, as shown in Figure 1(a), the lapse-rate tropopause is 0.33 km above the tropopause derived from curve fitting, and the transition depth is 0.82 km. However, for a weak transition profile, the difference and transition depth are much larger (0.80 and 2.78 km, respectively). This distinction between sharp and slow transition has been stated by Pan et al. (2004); the sharper the tropopause, the smaller the difference among various definitions of the tropopause and the narrower the transition depth.

3. Results

3.1. Tropopause-based mean structure and TIL

We begin by examining four individual stations selected to represent four different latitudinal bands; namely, Nenjiang (49.10°N), Beijing (39.48°N), Nanyang (33.02°N), and Huaihua (27.33°N). Figure 2 shows their tropopause-based mean seasonal profiles of temperature and \(N^2\) in spring, summer, autumn, and winter, respectively. Obviously, the tropopause altitude at all these stations has significant seasonal variation, particularly for higher latitude stations; for example, the seasonal mean tropopause altitude in Nenjiang is 8.7 km in winter, but increases to 11.8 km in summer. This feature, with a higher tropopause in summer and lower tropopause in winter, can be attributed to the annual variation in Brewer–Dobson circulation, which is stronger in winter and weaker in summer (Gettelman et al. 2011). The tropopause altitude also has a significant meridional distribution—higher at lower latitudes and lower at higher latitudes. At lower latitudes, such as Huaihua, the tropopause is often tropical in its features, whereas stations at higher latitudes (such as Nenjiang) are dominated by a polar tropopause (Bian 2009; Gettelman et al. 2011).

Just above the tropopause, an obvious TIL exists at all four stations, in which the temperature increases rapidly and gives rise to a sharp maximum of \(N^2\) (6.0 × 10⁻⁴ s⁻² to 8.0 × 10⁻⁴ s⁻²), and then \(N^2\) falls back slowly to typical stratospheric values (4.0 × 10⁻⁴ s⁻²). At higher latitudes

Figure 1. Two typical profiles of temperature (left-hand graph in each panel) and static stability (right-hand graph in each panel), and corresponding hyperbolic tangent fitting curves (red): (a) case with a sharp stability transition from troposphere to stratosphere; (b) case with slow transition.

Note: The tropopause levels, based on the WMO definition (black) and curve-fitting (red), are shown as transverse lines, respectively.
The layer with enhanced $N_2$ just above the tropopause is the TIL, above which the $N_2$ gradually returns to typical stratospheric values. It should be noted, however, that lower latitudes seem to have a thinner TIL and much higher typical stratospheric values of $N_2$.

In all seasons, the mean tropopause altitude drops quickly from low to high latitudes. The mean tropopause level is always above 16 km, and shows very little change at lower latitudes, which is dominated by the tropical tropopause and is higher. On the contrary, higher latitudes are dominated by a much lower polar tropopause. In the intermediate zone, the mean level of the tropopause is determined by the partition of the tropical tropopause versus the polar tropopause, which is related to the subtropical westerly jet. So, the meridional distribution of the tropopause exhibits an abrupt transition across the subtropical jet, which can be attributed to the occurrence of a tropopause fold in this region (Randel, Seidel, and Pan 2007).

3.2. TTL

The TTL is analyzed by the method described in Section 2.4. In order to demonstrate the impact of transition sharpness on identifying the level of the tropopause, Figure 4 shows a scatter plot of the curve-fitting tropopause versus the lapse-rate tropopause for all the soundings in Xilinhaote.

(Nenjiang, 49.10°N), the TIL shows a distinct seasonal variation in maximum values of $N_2$, being a bit larger in spring–summer ($7.9 \times 10^{-4}$ s$^{-2}$) than in autumn–winter ($6.1 \times 10^{-4}$ s$^{-2}$), and the thickness of the TIL is much larger in winter (2.9 km) than in spring–summer (1.3 km). The thickness of the TIL increases poleward: the annual mean thickness of the TIL is less than 1.0 km at lower latitudes (Nanyang, 33.02°N; Huaihua, 27.33°N), and increases to larger values (about 2 km) at higher latitudes. Taking Nenjiang (49.10°N) as an example, the annual mean thickness of the TIL is 2.7 km, which is close to the value for Munich, Germany, located at 48°N but with a quite different longitude (Birner, Dornbrack, and Schumann 2002).

A thicker TIL at higher latitudes and a thinner TIL at lower latitudes can be attributed to cyclonic/anticyclonic flow in the upper troposphere, according to Wirth (2003) and Randel, Wu, and Forster (2007), or to the radiative effects of ozone and water vapor in the upper troposphere and lower stratosphere, according to Birner (2006). This issue is still under investigation.

Figure 3 shows cross sections of seasonal tropopause-based mean profiles of temperature and squared buoyancy frequency. It notes: [(a–d) correspond to spring, summer, autumn and winter, respectively] at four stations: Nenjiang (black); Beijing (blue); Nanyang (red); and Huaihua (green). Horizontal lines denote the seasonal mean tropopause altitude.
indicated in cooler colors. However, when the TTL is broad (with larger $\lambda$, in warmer colors), the curve-fitting tropopause has a much larger negatively biased altitude relative to the lapse-rate tropopause, and has a more disperse distribution from the 1:1 line. Lower latitude stations (Yichang, as shown in Figure 4(b)) have a larger depth of transition layer and also a higher tropopause, whereas the opposite is true for higher latitude stations (Xilinhaote, Figure 4(a)), located to the north of the subtropical jet in all seasons.

Figure 3. Seasonal tropopause-based mean squared buoyancy frequency as a function of latitude and altitude: (a) spring; (b) summer; (c) autumn; (d) winter. Note: White lines denote the seasonal mean tropopause altitude.

Figure 4. Scatter plot of the curve-fitting tropopause versus the lapse-rate tropopause for all the radiosonde data from (a) Xilintaote and (b) Yichang. Note: The 1:1 line is plotted in black, and the transition depth is shown in color.

and Yichang, as two examples. In general, the two definitions agree very well for both stations, particularly for tropopause altitudes less than 14 km, which is considered the division level between the tropical tropopause and polar tropopause (Pan et al. 2004). On average, the curve-fitting tropopause is 0.3–0.5 km lower than the lapse-rate tropopause. If the TTL depth is taken into consideration, a much smaller discrepancy is seen between the two tropopauses, with different definitions for the sharp transition layer with smaller $\lambda$, as indicated in cooler colors. However, when the TTL is broad (with larger $\lambda$ in warmer colors), the curve-fitting tropopause has a much larger negatively biased altitude relative to the lapse-rate tropopause, and has a more disperse distribution from the 1:1 line. Lower latitude stations (Yichang, as shown in Figure 4(b)) have a larger depth of transition layer and also a higher tropopause, whereas the opposite is true for higher latitude stations (Xilinhaote, Figure 4(a)), located to the north of the subtropical jet in all seasons.
In order to clearly show the dependence of the TTL on the latitude, the eight stations are grouped into four latitudinal bands at intervals of 7°. Figure 5 shows the occurrence fraction of transition depth $\lambda$ in different latitudinal bands. Obviously, higher-latitude bands have a much larger fraction of sharp transitions, while lower latitude bands have a larger fraction of slow transitions (Figure 5(a)). For example, the 43°–50°N band has an annual fraction of 69.3% for $\lambda < 1.0$ km, whereas the 22°–29°N band has a fraction of 43.5% for $\lambda > 2.0$ km. If the annual fraction for $\lambda > 2.0$ km is calculated for each band, a decreasing trend can be seen with increasing latitude. As shown in Table 1, the annual fraction of broad transitions is 43.8% in Nanning, and it falls to 9.4% in Nenjiang. These results are similar to those of Homeyer, Bowman, and Pan (2010), who focused on continental United States in late spring and early summer. The dominance of thick transition layers in lower latitudes can be attributed to the dynamical and radiative characteristics of the tropical upper troposphere (Fueglistaler et al. 2009), but it does not explain the presence of all large values of $\lambda$ that occur at latitudes north of the subtropical jet. Bethan, Vaughan, and Reid (1996) suggested that deep transitions in midlatitudes primarily occur in cyclonic flows in the lower stratosphere: since potential vorticity is conserved, increasing cyclonic vorticity must be balanced by stretching the column; as the column stretches, the lapse-rate increases, and the transition layer becomes broader.

Next, we examine the seasonal variations of the transition depth distribution. Springtime and wintertime appear to have a much sharper tropopause, particularly at higher
latitudes (Figure 5(b) and (e)). The fractions for $\lambda < 0.5$ km are all above 45% in both the 43°–50°N and 36°–43°N band in spring and winter. One exception is the southernmost band, which also has a similarly large sharp tropopause fraction, whilst the 29°–36°N band has fewer sharp transitions but more deep transitions, which is opposite to the tendency of the occurrence of transition sharpness with latitude. This is probably related to the location of the subtropical jet in these seasons, which is right between these two bands. As stated by Wirth (2003) and Randel, Wu, and Forster (2007), the jump in static stability at the tropopause is significantly weaker in the presence of cyclonic flow in the upper troposphere. The band to the north of the jet has relatively higher positive vorticity, and therefore a deeper transition layer; and vice versa for the land to the south of the jet.

The thickness of the TTL becomes deeper in the summertime, especially at lower latitudes, as shown in Figure 5(c). The occurrence of deep transition ($\lambda > 2.0$ km) even exceeds 65% for the southernmost band, which is located in the ITCZ and naturally has the characteristics of the tropical upper troposphere.

4. Conclusion

In this study, the meridional thermal structure of the tropopause layer is investigated using high vertical resolution radiosonde measurements over China. First, from the tropopause-based mean method, the strong TIL is analyzed. The results show that a TIL exists over every station from north to south in each season. Higher latitudes have a much larger TIL thickness (~2 km) than lower latitudes (with a TIL thickness less than 1.0 km). However, the formation mechanism is still an ongoing issue.

The structure of static stability near the tropopause is analyzed by fitting a hyperbolic tangent function to radiosonde profiles of Brunt–Vaisala frequency $N(z)$. The fitting parameters include the location of the tropopause ($z_0$) and the depth of the transition layer ($\lambda$). The eight radiosonde stations are grouped into four latitudinal zones at intervals of 7°. According to the value of $\lambda$, the tropopause transition is divided into sharp and deep transition ($\lambda < 2$ km or $\lambda > 2$ km). With increasing latitude, the fraction of profiles with a deep transition layer decreases rapidly. The tropopause at higher latitudes is much more likely to be sharp and have a TIL. The fraction distributions of $\lambda$ have seasonal variation both for lower latitudes and higher latitudes. The dominance of thick transition layers at lower latitudes can be attributed to the characteristics of the tropical upper troposphere (Fueglistaler et al. 2009) and the strong vertical gradient of the downwelling branch of the residual circulation just above the tropopause (Birner 2010), but other factors such as vorticity, the subtropical jet, and location of the ITCZ have an impact on the detailed distribution of the transition layer depth.

It should be noted that the length of the data used in this work is less than three years, which may cause some uncertainty in the results. Obviously, these data cannot reveal the interannual or decadal variations, which would require a much longer data period. Since the structure of the tropopause has significant variations both seasonally and meridionally, the results derived from these data of short length is still reasonable, and can provide some insight for further study.

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