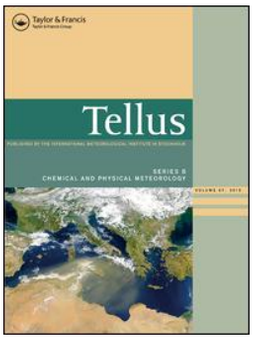

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Effects of the Tibetan Plateau on total column ozone distribution

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ABSTRACT

The relatively low total column ozone (TCO) above the Tibetan Plateau (TP) observed in summer is only partly due to the thinness of the atmospheric column. In this paper the effect of the TP on the TCO is further investigated using satellite data [Total Ozone Mapping Spectrometer (TOMS) ozone column and Stratospheric Aerosol and Gas Experiment II (SAGE II) ozone profiles], ECMWF ERA-40 reanalysis data and a 3-D chemistry-climate model (CCM). It is found that the low TCO over the TP is also closely related to large-scale uplift and descent of isentropic surfaces implied by seasonal and longitudinal variations in the tropopause height. The variations in tropopause height, with a maximum in summer, can be driven by various processes including convective activity, air expansion as well as the monsoon system. While previous studies have showed an important role of troposphere-to-stratosphere transport in contributing to the observed low ozone column over the TP, the mechanism revealed in this study is an alternative amendment to the causes of the TCO low over the TP. It is also found that the monsoon anticyclone circulation induces an isentropic transport of trace gases from high latitudes towards the TP in the lower stratosphere and hence modifies tracer distributions. For the vertical distribution of ozone, the modulation by the TP is most significant below ∼20 km, that is, in the upper troposphere and lower stratosphere (UTLS). The smaller differences in NO_x between Eastern TP and TP compared to large dynamically caused differences in ozone and methane imply the TCO low over the TP is mainly due to transport processes rather than chemistry.

1. Introduction

In the northern hemisphere mid-latitudes the Earth's surface is significantly uneven. The Tibetan Plateau (TP) is the highest landmass in the world, located in the latitude band 27.5°–37.5°N with an average height of more than 4000 m above sea level. Previous studies have revealed the so-called 'ozone valley', monthly average total column ozone (TCO) over the TP about 20–30 Dobson Unit (DU) lower than other regions at similar latitudes from June to September. These have argued that this low column ozone is caused mainly by upward motions over the mountain region, which carry ozone-poor air from the troposphere into the lower stratosphere (Zhou et al., 1995; Zou, 1996; Zheng et al., 2004; Zhou and Zhang, 2005). Hingane (1990) also found that a regular, quasi-stationary ozone minimum region exists over the Himalayas and northern India between spring and summer. This low column ozone region may occasionally drift northwards, forming ozone mini-holes away from its original region (James, 1998).

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There is no doubt that the TP has important thermal and dynamical impacts on the general circulation above it. However, to what extent the so-called 'ozone valley' is affected by the transport processes associated with upward motion over the mountain region is not well understood. Due to the high altitude of the TP, the surface air density is about 60–70% of that at sea level. It has been found that the absence of about 4 km air column itself can lead to a reduction of TCO by around 2.5%, but this is not enough to account for all the observed decrease of column ozone over the TP in summer, which may reach 11% in May. Coldewey-Egbers et al. (2005) studied the effect of the Himalayas on satellite-retrieved ozone column and found that a reduction of about 10 DU in the total column can be caused at a terrain height of 4 km, but the observed 'ozone valley' is about 20–30 DU lower. Previous studies suggested that the thermal-dynamical forcing of the TP, for example, forced ascent, thermal convection, monsoon system, air expansion, makes a significant contribution to the TCO low (e.g. Ye and Xu, 2003; Randel and Park, 2006). However, how and to what extent the thermal-dynamical forcing of the TP modify TCO over the TP are still a subject of much debate.

Zhou et al. (1995) suggested that the total column low over the TP is mainly caused by upward motion over the

mountain region with chemical processes playing a secondary role. However, upward motions over the mountain region are mostly confined in the troposphere and air in the upper troposphere and lower stratosphere (UTLS) generally tends to 'flow over' mountains at relatively high speeds, that is, high mountains can cause an upward displacement of streamlines rather than taking tropospheric air irreversibly into the stratosphere. Even though the thermally induced deep convective systems can occasionally bring tropospheric air into the stratosphere, the net ozone flux across tropopause is downward in the long-term mean (e.g. Dethof et al., 2000; Collins et al., 2003). Ye and Xu (2003) proposed that a negative, vertical advection of ozone concentration and air expansion associated with the thermally forced local circulation contribute partly to the total column low over the TP. It should be noted that in the troposphere, ozone concentrations are generally small and the modulation of the tropospheric circulation on the TCO is not, on average, significant. The implied mechanism in these previous studies is that the terrain-induced local circulations which modify TCO are mainly related to stratosphere–troposphere exchange (STE) processes.

Randel and Park (2006) showed that the Asian summer monsoon anticyclone has strong influences on constituent behaviour in the upper troposphere through the coupling of circulation and convection. Their idealised model results indicate that ozone and water vapour anomalies are confined within the upper tropospheric anticyclone and imply that the Asian summer monsoon anticyclone may also affect the TCO low over the TP. However, Jackson et al. (1998) pointed out that highest water vapour concentrations in the lower stratosphere occur over the TP rather than over the monsoon region and Fu et al. (2006) showed that tropospheric deep convection driven by elevated heating above TP is responsible for the cross-tropopause transport of water vapour over the TP.

Moore and Semple (2005) pointed out that a Taylor Cap exists over the TP which implies that the Tibetan Plateau is exerting an influence into the lower stratosphere resulting in a halo of high surrounding TCO. Some other observational and modelling studies, however, argued that high ozone levels around the TP are mainly related to STE (Ding and Wang, 2006) or regional transport of polluted air from boundary layer (Zhu et al., 2004). A question arises whether elevated ozone levels around the TP are also a result of the TCO low over the TP. On the other hand, the role of chemical processes in contributing to the low TCO is also a question of doubt. From satellite observations Fishman et al. (2003) found evidence of high tropospheric ozone abundances in the eastern TP regions while previous modelling studies showed that tropospheric chemical and transport processes may result in high ozone values in eastern Asia (e.g. Zhu et al., 2004), However, the contribution of the chemical processes in the lower stratosphere to the TCO low over the TP is still not clear.

In this paper, we combine various data sets to investigate the thermal-dynamical effects of the TP on TCO distribution in an attempt to further clarify the main causes of the TCO low over the TP, and the effect of UTLS chemical processes on the TCO low is analysed using CCM simulations. Section 2 gives a brief description of the data used, Section 3 presents the TCO and tropopause height analysis. Section 4 discusses the possible mechanisms responsible for the TCO low and Section 5 gives our summary.

2. Data description

Monthly mean Total Ozone Mapping Spectrometer (TOMS) data from 1979–1992 with a resolution of 1° latitude × 1.25° longitude was obtained from the TOMS website (ftp://toms.gsfc.nasa.gov/pub/eptoms/data/). Stratospheric Aerosol and Gas Experiment II (SAGE II) observations (version 6.20) are used to analyse vertical profiles of ozone concentration. The European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-40 reanalysis data from 1979 to 1992 are used to derive the tropopause height. For the purpose of diagnosing the role of chemical and transport processes in modulating ozone concentrations in the lower stratosphere, 20-yr chemistry-climate model (CCM) output from the UK Met Office Unified Model (UM) coupled with a detailed chemistry scheme (Chipperfield, 1999) is also used. Details of the CCM can be found in Tian and Chipperfield (2005). The CCM advects 28 tracers with around 42 chemical species including O_x, HO_x, Cl_y, Br_y and NO_y families. The model includes both gas-phase chemistry and heterogeneous chemistry on liquid and solid aerosols. The CCM data has a horizontal resolution of 2.5° × 3.75° and 64 levels from the surface to 0.01 hPa with a vertical resolution of about 1.5 km in the UTLS region. The chemistry is calculated on 30 levels starting from ∼150 to 0.5 hPa. A 20-yr transient run, with the observed sea surface temperatures (SSTs) and sea ice from the Atmospheric Model Intercomparison Project II (AMIPII), was performed. The CCM performance and some validation of this 20-yr CCM data set can be found in Eyring et al. (2006).

3. TCO low and tropopause height over the TP

Figure 1a shows the longitude-time section of the TOMS TCO climatology (1982–1992) averaged within the 27.5°–37.5°N latitude band. Also shown is the corresponding 20-yr TCO climatology from the CCM output (Fig. 1b). Note that the CCM calculates chemistry at 150 hPa and above and a tropospheric zonal mean ozone climatology from Logan (1999) has been added to the modelled TCO. Hereafter, all CCM fields shown are 20-yr averaged climatologies unless otherwise stated.

Consistent with the previous findings, the column ozone over the TP (75°–105°E) is generally lower than over the surrounding longitudes. The differences are more significant from April to August. We can note that the TOMS TCO values over the North American Rockies (from 239° to 260°E) are also lower than in the vicinity, although this feature is less significant than that over the TP. Similar features can be seen in the corresponding CCM

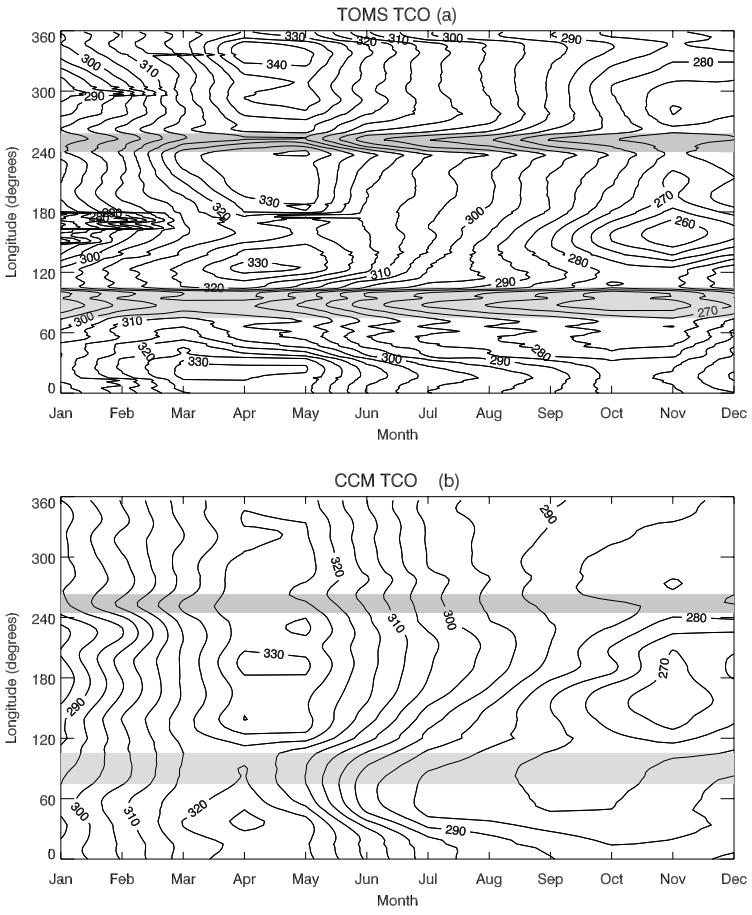

Fig. 1. (a) Longitude-time cross-section of total column ozone (TCO, DU) averaged over the 27.5°–37.5°N latitude band from (a) TOMS (1982–1992) and (b) the CCM (20-yr mean). The locations of the Tibetan Plateau (TP) and Rockies are shaded light grey and grey, respectively.

TCO field (Fig. 1b). However, due to the addition of a longitudinally constant tropospheric contribution and the smoothed topography at the model's low spatial resolution, the TCO low is less significant than in the TOMS TCO field. To illustrate this, Fig. 2a shows the tropospheric ozone residual (TOR) as described by Fishman et al. (2003). The CCM column ozone averaged over the 27.5°–37.5°N latitude band with the Logan (1999) tropospheric ozone climatology replaced by the TOR is also shown in Fig. 2b. Note that the TOR is significantly lower over the TP and Rockies than their surrounding longitudes, reflecting the contribution of the thinner air column over these elevated land surfaces. When the TOR is added to the CCM ozone column, the TCO low over the TP and Rockies is as significant as that in the TOMS data (Fig. 1a).

Figures 1 and 2 clearly show that the Earth's topography has an impact on the TCO distribution and the impact can be noted in both the tropospheric and stratospheric ozone column. The TCO is generally smaller over continents than over oceans. Previous studies have pointed out that differences in the thickness of the air column between mountain and non-mountain regions can account for part of the total column reduction over the TP (e.g. Ye and Xu, 2003). Air columns over the TP and the Rockies are significantly thinner than over oceans; therefore, persistent low TCO values can be observed over those regions. A key question addressed in this study is then what is the role of other processes in modulating the TCO over the TP.

One may expect that the high topography tends to lift the streamlines upward over the TP. The upward lifting of isentropic surfaces will result in lower ozone concentrations at a given altitude over the TP compared to that over flat regions. To clarify this argument, it is necessary to examine the tropopause variation over the TP. Figure 3 shows the longitude-time sections of the thermal and the cold-point tropopause heights (averaged between 27.5° and 37.5°N) derived from ERA-40 data. The thermal tropopause height follows the World Meteorological Organization (WMO) definition and the cold-point tropopause height is defined as the altitude of the temperature minimum. It has been pointed out that the thermal tropopause has limited physical relevance especially in the tropics while the cold-point tropopause correlates better with convective processes which may play an important role in the STE (e.g. Highwood, 1998; Schmidt et al., 2004).

There are significant differences in the altitude of the tropopause calculated by the two methods. Fig. 3 indicates that cold-point tropopause heights over the TP are higher than thermal tropopause heights in the same region. Schmidt et al. (2004)

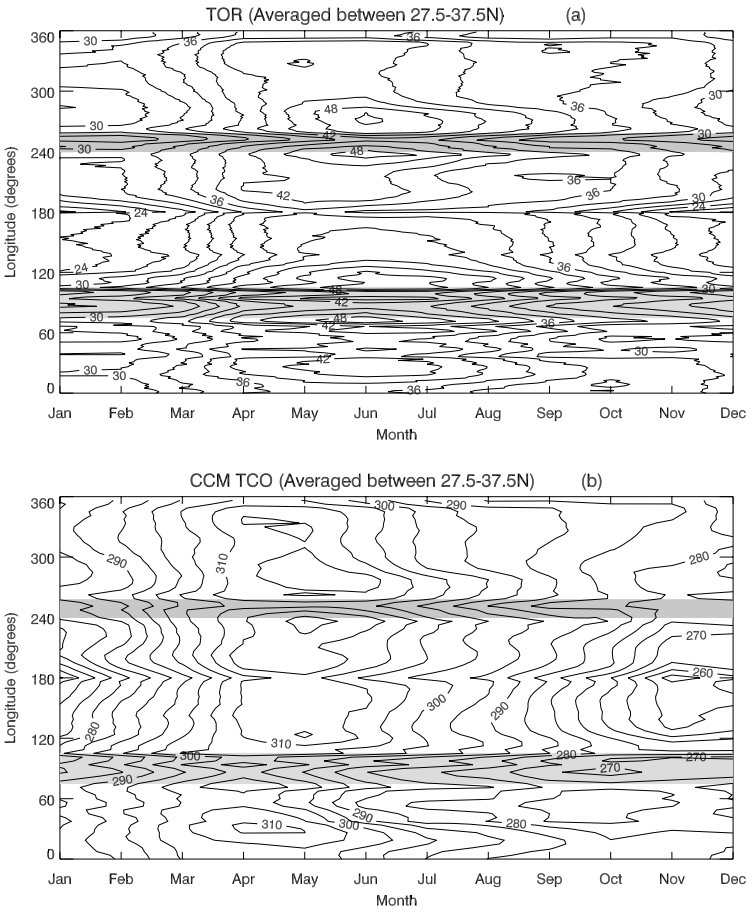

Fig. 2. (a) Longitude-time cross-section of the tropospheric ozone residual (Fishman et al., 2003) averaged over the 27.5°–37.5°N latitude band. (b) the CCM (20-yr mean) column ozone averaged over the 27.5°–37.5°N latitude band with the corresponding TOR replacing the Logan (1999) climatology in the troposphere. The locations of the Tibetan Plateau (TP) and Rockies are shaded light grey and grey, respectively.

also found, from GPS radio occultation measurements, that the cold-point tropopause height is generally higher than the thermal tropopause. It is interesting that the thermal tropopause heights over the TP reach a maximum (of nearly 16 km) in July and August while the cold-point tropopause reaches a maximum (of ∼17 km) earlier in May. One possible reason is that the Southeast Asian monsoon usually occurs in May (He et al., 1987) which implies more convective processes and hence higher cold-point tropopause heights. In late summer, a thermal-dynamical equilibrium may be reached over the TP meaning that the thermal tropopause reaches its maximum in July and August. Overall, tropopause heights over the TP are higher than elsewhere in summer and a similar behaviour can be noted over the Rockies in North America.

Figure 4 shows the corresponding tropopause heights derived from the CCM output. The patterns in the CCM tropopause heights are generally consistent with those in the ERA-40 data. Also notable is that the cold-point tropopause height over the TP in the CCM reaches a maximum in June, one month later than in the ERA-40 data, possibly because the onset of the South Asia monsoon is not well captured in our CCM. By looking at Figs. 1–4 together, we can see that variations in the tropopause height have a close relation to the TCO distribution over the TP with higher (lower) tropopause heights corresponding to low (high) TCO values.

Figure 5 shows the differences in the TOMS and CCM TCO between the TP (70°–105°E, 27.5°–37.5°N) and the eastern TP (ETP, 105°–135°E, 27.5°–37.5°N). Also shown are the corresponding differences in the ERA-40 cold-point and thermal tropopause heights (Fig. 5c). To estimate the contribution of the thinness of air column to the TCO low, the reference climatology of ozone profiles from Fortuin and Kelder (1998) was used together with topography data. The contribution of the thinness of air column at any selected point is taken as the integral of the reference ozone profile from 0 km to the elevation height of this point. Fig. 5a also gives the differences in estimated partial column ozone due to the thinness of air column between the above two selected regions. It should be pointed out that the ozone profile climatology of Fortuin and Kelder (1998) has uncertainties particularly over the TP due to the sparseness of measurements in this region. Satellite observations are available, but coarse vertical resolution in the lower troposphere prevents us from using them to estimate the TCO reduction caused by the thin air column. On the other hand, very limited sounding profiles are available over the TP, therefore, the ozone profile climatology of Fortuin and Kelder (1998) is used here as a substitute in

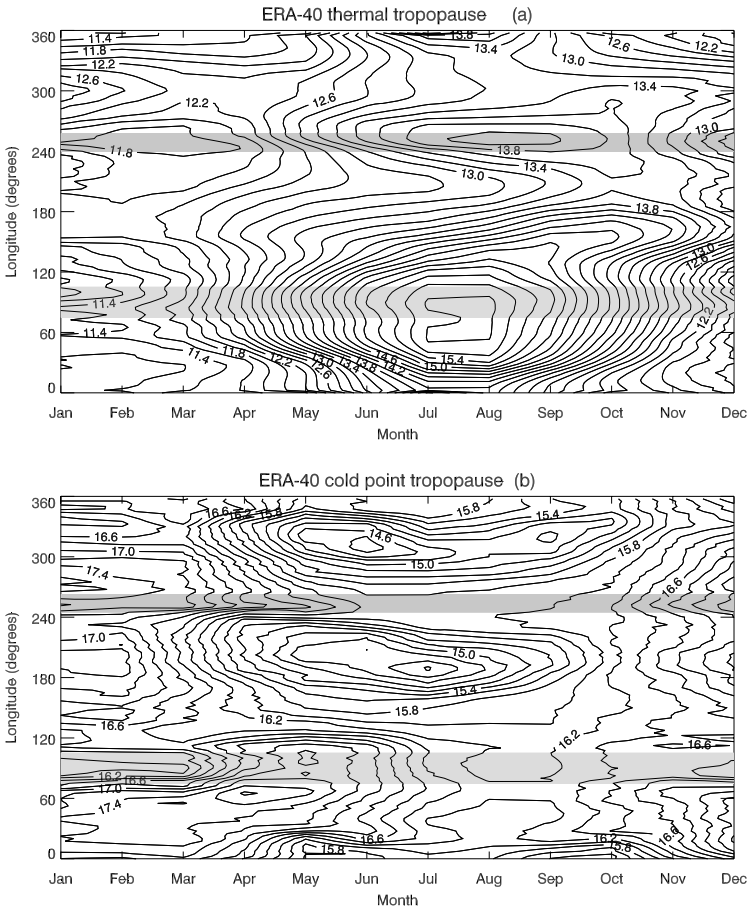

Fig. 3. Longitude-time cross-sections of (a) the cold-point and (b) thermal tropopause heights (km) derived from ERA-40 data (1982–1992) averaged over the 27.5°–37.5°N latitude band. Contour interval is 0.2 km. The locations of the TP and Rockies are shaded light grey and grey, respectively.

our estimation. Fig. 5a indicates that the estimated ozone column reduction due to the elevated height of the TP is about 10–14 DU. Based on the estimation from the ozone soundings over Lasa (Ye and Xu, 2003) and satellite-retrieved ozone columns (Coldewey-Egbers et al., 2005), the absence of 4 km air column has previously been estimated to result in a reduction of TCO of no more than 10 DU in August (about 2.4% of the TCO). Our estimation is slightly higher, but nevertheless close to it.

Figure 5a indicates that the largest TOMS TCO differences between the TP and the ETP occur in April, May and June. Accordingly, the ERA-40 cold-point and thermal tropopause height differences maximise in May and June, respectively. In contrast, the largest CCM TCO differences occur in May, June and July, one month later than in the TOMS data. The timing is in accordance with the occurrence of the largest cold-point tropopause heights in Figs. 3 and 4, but not for the thermal tropopause height which is highest in July and August. Also note that the CCM TCO differences are only about half of those of the TOMS TCO. This is due to the absence of the tropospheric contribution in the CCM chemistry which is calculated from 150 hPa and above. As a result, the tropospheric contribution to the TCO low is excluded in the CCM calculation. From another point of view, the smaller CCM TCO differences implies that both stratospheric and tropospheric ozone contribute to the TCO low over the TP. From April to August the reduction in TCO due to the thin air column reaches 60% of the TOMS differences. As a further examination of the above estimate, Fig. 5b gives the TOR and the TOMS column ozone differences between the TP and the ETP. Here, the contribution of the thinner air column estimated from climatological ozone profiles is excluded from the TOMS TCO. From Fig. 5b, we can see that the stratospheric contribution to the TCO low over the TP, as implied by the CCM column ozone differences, is close to that calculated from the TOMS column ozone with the terrain's contribution excluded. Note that the TOR differences in Fig. 5b reflect the tropospheric contribution to the TCO low over the TP. Figs. 5a and b suggest that a contribution of about 10 DU to the TCO low is from the stratosphere, while the troposphere contribution is around 5 DU.

The magnitude of the TCO low is dependent on the region selected for comparison. If the TCO over the western TP (WTP, 45°–75°E, 27.5°–37.5°N) is considered, the corresponding values in Fig. 5a become smaller (not shown). One reason is that the differences in the topography and tropopause heights between the TP and the WTP are relatively small. Another reason

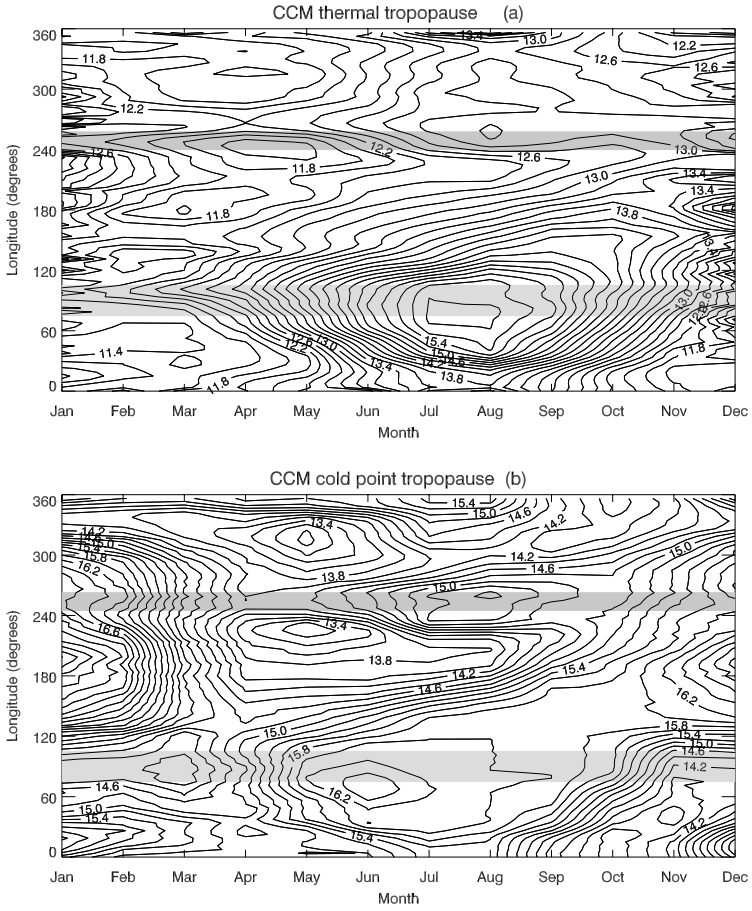

Fig. 4. Longitude-time cross-sections of (a) the cold-point and (b) thermal tropopause heights derived from the CCM output (20-yr means) averaged over the 27.5°–37.5°N latitude band. Contour interval is 0.2 km. The locations of the TP and Rockies are shaded light grey and grey, respectively.

is that the tropospheric ozone abundances over the ETP are larger than over the WTP due to the air pollution in the eastern Asia (Fishman et al., 2003). On the other hand, satellite observations have revealed that deep moist convection frequently occurs over the ETP (Chen and Liu, 2005) implying ozone profiles are more likely be affected by this process over this region. If the TCO is averaged between 0°–360°E, but with the exclusion of the band 75°–105°E (the same definition for the non-mountain region in Zhou and Zhang, 2005), the features are similar to Fig. 5 but the magnitudes are even larger than corresponding values for the ETP. We can also note from Fig. 5c that the differences in the tropopause height between the TP and ETP are much smaller than those between the TP and the non-mountain region. This is because the differences between the TP and non-mountain region result from the combined effects of topography and ocean surface. It should be pointed out that the CCM TCO over the TP in winter months is even larger than surrounding areas while the TOMS TCO over the TP is still lower than its vicinity. As will be shown later, the CCM tropopause in DJF is significantly lower than the ERA-40 tropopause over the TP. This may be the main reason for too large ozone over the TP in the CCM for winter.

4. Thermal and dynamical effects of the TP on TCO low

A high orography such as the TP tends to lift streamlines upward, implying a higher tropopause. On the other hand, the high summertime tropopause is due to the expansion of air caused by thermal forcing as the TP acts as an elevated heat source in summer. Additionally, the Asia summer monsoon systems and the thermal effect of the TP will result in more convective activity in summer. This convective activity also tends to increase average tropopause height over the TP. The so-called Tibetan High in the upper troposphere over the TP (e.g. Murakami, 1987), which has been mentioned as one of the most important factors for the generation and maintenance of Indian summer monsoon, may also tend to maintain a high tropopause over the TP in summer. In winter, however, as the TP becomes a cold thermal sink, the whole air column tends to shrink and prevailing descent of flow over the TP can occur (e.g. Yanai et al., 1992). Consequently, the tropopause over the TP is significantly lower than the surrounding area.

Previous studies have argued that local circulations induced by the thermal and dynamical forcing of the TP can move ozone-rich (poor) air downward (upward) and hence cause an increase

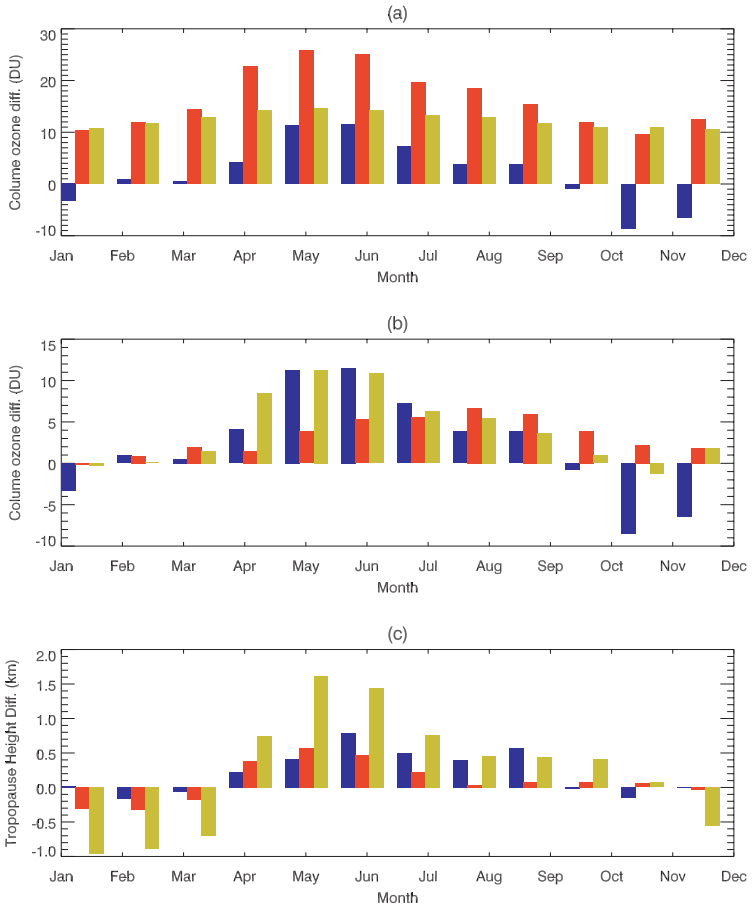

Fig. 5. (a) The difference in TCO (DU) between the TP (75°–105°E) and the ETP (105°–135°E) for TOMS data (red) and the CCM (blue). Also shown is the partial column ozone difference due to the thin air column between those two regions (light yellow) estimated from climatological ozone observations (see text for more details). (b) The corresponding differences in TOR (red) and TOMS ozone excluding the tropospheric contribution estimated from climatological ozone observations (light yellow) between the TP and the ETP. The corresponding difference in CCM column ozone (no tropospheric contribution) from panel (a) are replotted for reference (blue). (c) The differences in the ERA-40 tropopause heights (km) between the TP and the ETP for the cold-point (red) and thermal (blue) tropopause. The difference in the ERA-40 cold-point tropopause height between the TP and non-mountain regions (excluding 75°–105°E band) are shown in light yellow.

(decrease) of TCO (e.g. Zhou et al., 1995; Ye and Xu, 2003; Fu et al., 2006). It is important to distinguish here between the descent/ascent of isentropic surfaces in the UTLS from deep convective activities which are mainly confined in the troposphere with relatively short time scales. The integrated effect of terrain-induced deep convective activities may also act to maintain a higher tropopause height which implies an uplift of streamlines and upward shift of ozone vertical profiles.

Figure 6 shows the summer, winter and annual mean vertical profiles of ozone number densities over the TP (averaged over 75°–105°E, 27.5°–37.5°N from SAGE II observations for the period 1985–2003) and the corresponding longitudinal variation of tropopause heights derived from ERA-40 data and the CCM output. The average topography data within this latitude band [derived from the Global Land One-km Base Elevation (GLOBE) Project (1999)] are also shown for reference. As the SAGE II time series have missing values, a linear least-square regression method used by Zawodny and McCormick (1991) was used to fill in gaps. We can see that the cold-point tropopause height over the TP derived from ERA-40 data (Fig. 6b) is overall higher than elsewhere in summer (JJA mean) and lower in winter (DJF mean). In contrast, the tropopause height over the Pacific ocean is low in summer and high in winter. Correspondingly, the ozone vertical profile over the TP is shifted up in summer and down in winter (by about ±2–3 km) compared to annual mean profile. The higher tropopause in summer over the TP can cause a large-scale vertical movement of isentropic surfaces, and, consequently, an integrated shift of the ozone vertical profile. This upward shift of ozone profiles is one of the causes for the TCO low over the TP. Upward motion of air must be accompanied by mass divergence at high altitudes but the key point here is that at lower pressures a certain ozone mixing ratio, which is preserved by transport, will make a smaller contribution to the column. Integrating the three profiles in the vertical we get the column ozone values above 10 km, that is, 252, 267 and 258 DU for winter, summer and annual mean profiles, respectively. Comparing the summer and annual mean ozone profiles, a 6 DU reduction of the column ozone above 10 km can be caused due the upward shift of the summer ozone profile.

Figure 7 further shows JJA, DJF and annual mean CCM ozone and methane profiles for different regions. Note that the summertime ozone profile over the TP is clearly shifted upward in the lower stratosphere compared to winter and annual mean profiles (Fig. 7a). The upward shift of the summer ozone profile over the Pacific region (140°–180°E) is not significant relative to the winter and annual mean profiles (Fig. 7b). Fig. 7c indicates that

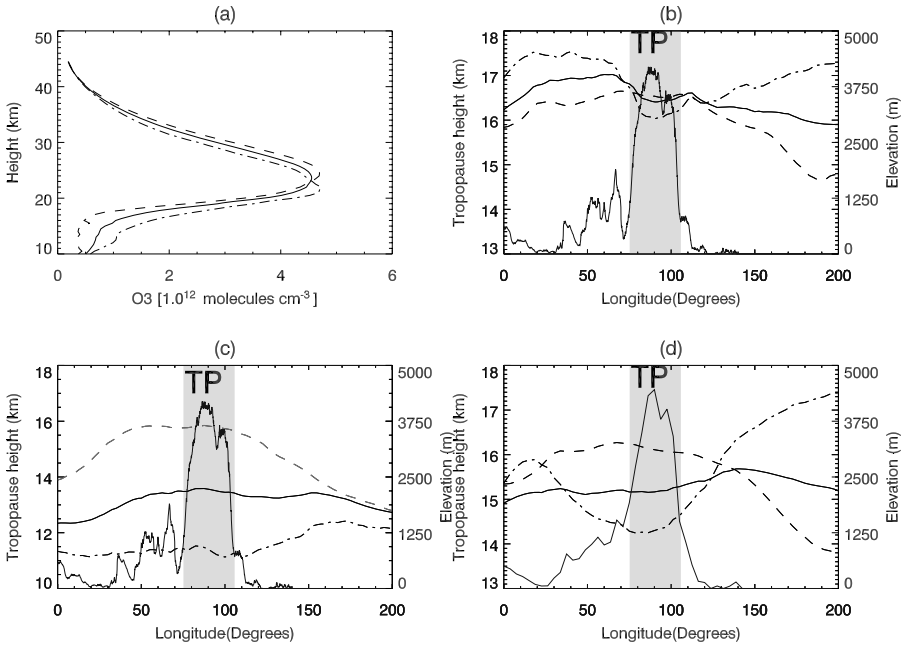

Fig. 6. (a) The summer, winter and annual mean vertical profiles of ozone number density over the TP (averaged over 75°–105°E, 27.5°–37.5°N) from SAGE II observations for the period 1985 to 2003 and (b) the corresponding longitudinal variation of the ERA-40 cold-point tropopause height averaged over the 27.5°–37.5°N latitude band. The dashed lines represent the summer (JJA mean) profiles, the dash–dotted lines represent the winter (DJF mean) profiles, and the solid lines represent the annual mean profiles. The topography data averaged over 27.5°–37.5°N are also shown for reference (heavy solid line, the right-hand y axis is for topography altitude). (c) As (b), but for the ERA-40 thermal tropopause height. (d) As (b) but for the CCM cold-point tropopause height and the corresponding model topography.

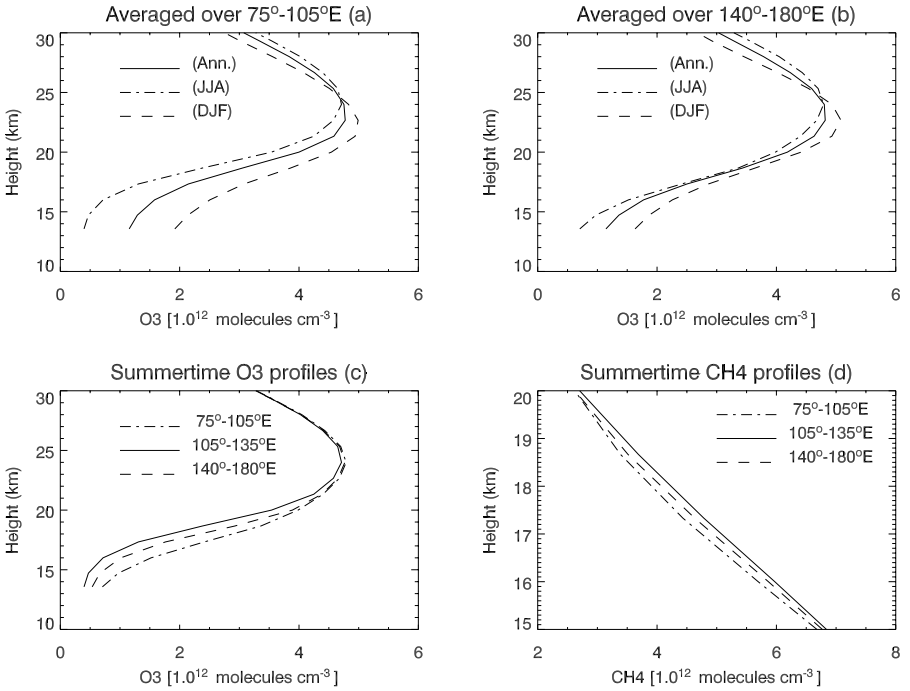

Fig. 7. (a) The summer (JJA, dash–dotted line), winter (DJF, dashed line) and annual mean (solid line) vertical profiles of the CCM ozone number densities over the TP (averaged over 75°–105°E, 27.5°–37.5°N, 20-yr mean). (b) As in (a) but for profiles averaged over 140°–180°E, 27.5°–37.5°N. (c) The summertime CCM ozone profiles averaged over the same latitude range (27.5°–37.5°N) but different longitude ranges, that is, 75°–105°E (dash–dotted line), 105°–135°E (solid line), 140°–180°E (dashed line). (d) As in (c) but for methane profiles.

the summertime ozone over the TP is lower than that over other regions in the lower stratosphere. However, we can note that differences between the profiles in Fig. 7 become small above 25 km. The result suggests that the upward shift of ozone profiles due to summertime tropopause variation is significant only in the lower stratosphere, that is, below 25 km. In contrast, and due to its opposite vertical gradient, methane concentrations over ETP and the Pacific region are lower than those over the TP in summer and higher in winter (Fig. 7d).

The column ozone values corresponding to the three ozone profiles in Fig. 7a are 262, 296 and 280 DU for winter, summer and annual mean ozone profiles, respectively. Comparing Figs. 6a and 7a, we can see that the overall features are the same between the modelled and SAGE II ozone profiles. However, the ozone column estimated from the CCM profiles is about 10 DU larger than that from SAGE II profiles, possibly due to coarse resolution and the approximate pressure-height conversion for the CCM profiles (the model pressure levels are converted to height levels approximately by $z = 16 \log(1000/p)$, where p is pressure in hPa, z is height in km).

Figure 8 gives the longitude-height cross-sections of the mean CCM potential temperature in summer. Note that upward shift of isentropic surfaces can be clearly seen in Fig. 8a. This upward shift of isentropic surfaces over the TP occurs mainly in the UTLS region. In the middle troposphere, a downward shift of isentropic surfaces over the TP is evident (Fig. 8b). These results confirm that the large-scale vertical shift of isentropic surfaces and accompanying shift of ozone vertical profiles over the TP partly contribute to the TCO low over the TP.

It has been found from observations that the thermally forced large-scale vertical tropospheric circulation over the TP is generally accompanied by the subsidence in the surrounding desert areas (He et al., 1987). Ye and Xu (2003) did not find anomalously high ozone concentrations in the descending regimes. They argued that the possible reason is that the rising area is much smaller than the sinking area so that downdraft velocities associated with positive transport of ozone are much smaller than those in updraft areas. It should be pointed that the accompanying subsidence mainly redistributes tropospheric air in which ozone concentrations are generally small. The thermally induced deep convective systems can occasionally bring tropospheric air into the stratosphere. However, those air masses of tropospheric origin will then be resident in the stratosphere for a long time, that is, months to years. Therefore, there should be no corresponding increases in ozone concentrations in the accompanying subsiding area except in events of intrusions of ozone-rich stratospheric

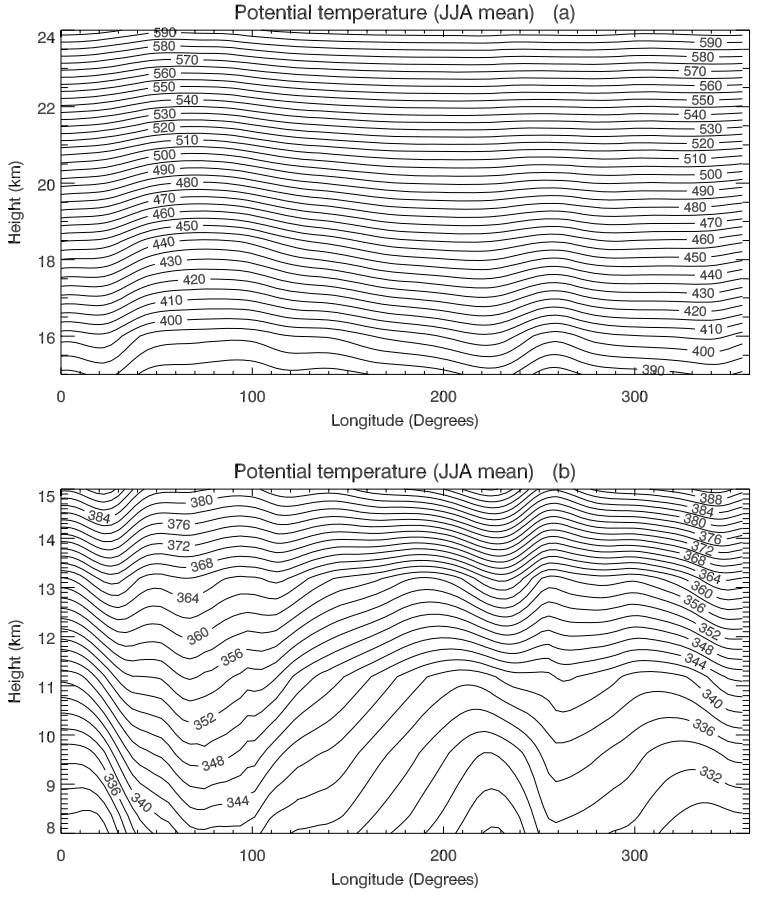

Fig. 8. Longitude-height cross-section of the CCM potential temperature (20-yr mean) averaged over the 27.5°–37.5°N latitude band for (a) 15–24 km (contour interval 5 K) and (b) 8–15 km (contour interval 2 K).

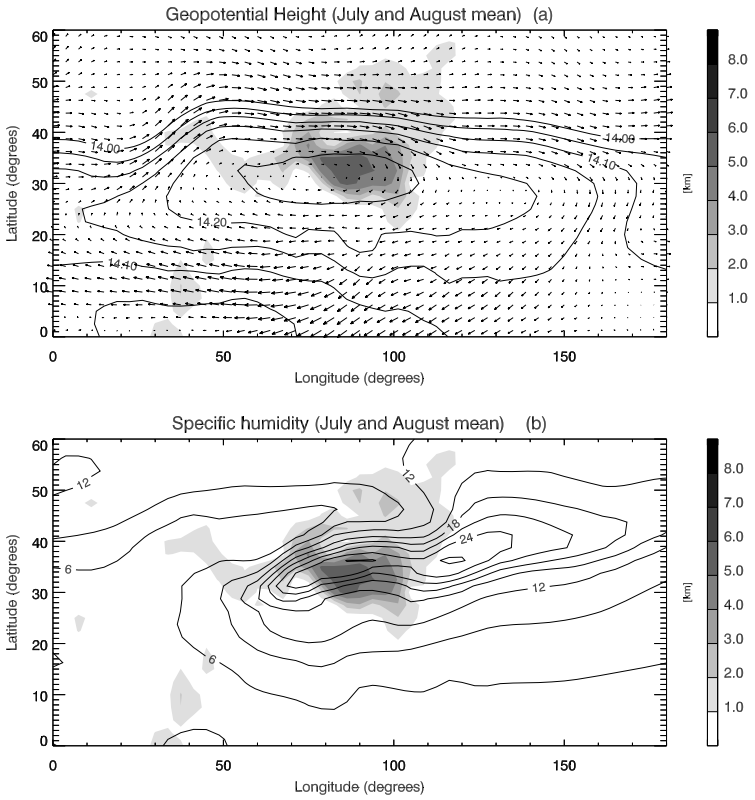

Fig. 9. Mean 20-yr CCM summer (July and August) latitude-longitude cross-sections on the 360 K isentropic surface for (a) geopotential height (contour interval 50 m) and horizontal winds (m s^{-1}), and (b) specific humidity (contour interval 3 ppmm). The filled contours represent topography (km).

air. The observed summertime maximum ozone at the northeastern boundary of the TP (e.g. Ding and Wang, 2006) may not be a compensation for the TCO low over the TP.

However, it appears that upward shift of ozone profiles over the TP in summer is not the only effect of convection on the TCO low. Although the convective activities due to the thermal effect of the TP are mostly confined within the troposphere, and the vertical development of Indian monsoon trough is thought to be restricted below 300 hPa by the Tibetan High (Chen and Yoon, 2000), Randel and Park (2006) found evidence that the Asian monsoon anticyclone affects tracer distributions in the UTLS region through its coupling with convection below. Fu et al. (2006) showed that the thermally induced convection over the TP can transport air masses from the upper troposphere to the lower stratosphere. In a similar approach to Randel and Park (2006), Fig. 9 shows the CCM July and August mean geopotential height and specific humidity field on the 360 K isentropic surface. Consistent with the result of Randel and Park (2006), the Tibetan High is clearly seen in the CCM geopotential and wind field and the water vapour is significantly higher within this anticyclone region. Also note that the larger CCM specific humidity values on the 360 K isentropic surface mainly occur over the ETP which is consistent with the finding that there is more deep moist convective activity in this region.

While the 360 K isentropic surface (around 12 km) over the TP is still in the troposphere in summer (see Fig. 4), one may wonder whether similar features can be found above the 360 K surface. Figure 10 shows the CCM ozone and methane fields on the 450 K isentropic surface (around 19 km). In previous studies (e.g. Hoskins, 1991) the 380 K surface has been regarded as the summer tropopause over the TP. Therefore, the 450 K surface should be well in the so-called overworld.

It is apparent that the Tibetan High and associated anticyclone have a significant effect on the ozone and methane distributions at this level. Fig. 10 indicates that there is ozone and methane transport from high latitudes towards the TP associated with the anticyclone circulation on the 450 K surface. As a result, there is larger ozone and lower methane located over the eastern TP. Comparing Fig. 9 with Fig. 10, We can see that the water vapour distribution on the 360 K surface is different from the ozone and methane distributions on the 450 K surface. The high water vapour over the TP on the 360 K surface is mainly related to the convective activities while the distributions of ozone and methane at 450 K is dominated by the isentropic transport associated with the anticyclone circulation over the TP. The STE processes appear to have no significant impact on ozone and methane distributions at 450 K. The above analysis suggests that apart from the upward shift of isentropic surfaces in the lower stratosphere due to high topography and tropopause, the anticyclone circulation-induced isentropic transport also plays a role in causing the TCO low over the TP. On the other hand, although the mechanisms proposed by previous studies (Randel

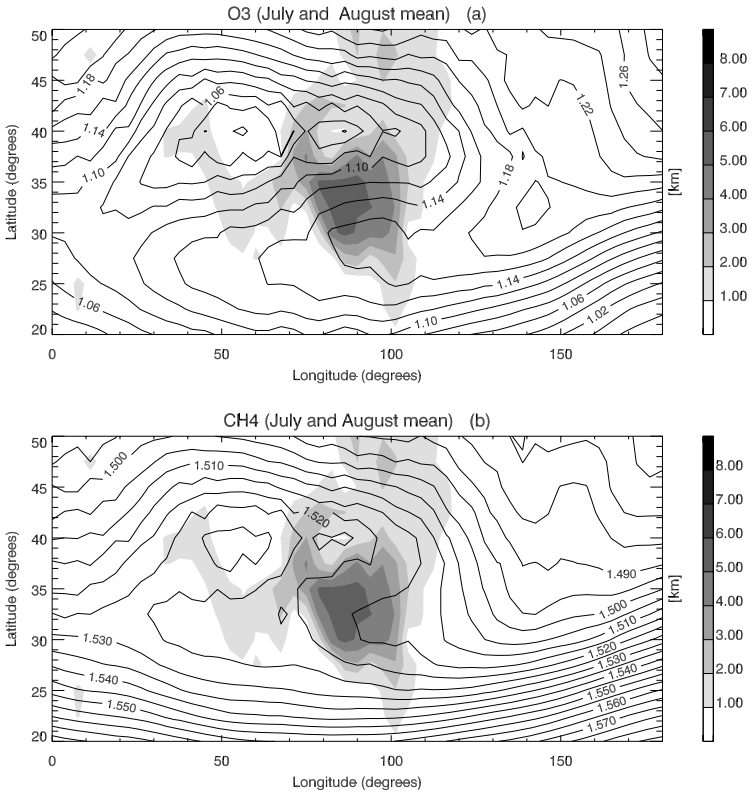

Fig. 10. Mean 20-yr CCM summer (July and August) latitude-longitude cross-sections on the 450 K isentropic surface for (a) ozone (contour interval 20 ppbv) and (b) methane (contour interval 5 ppbv). The filled contours represent topography (km).

and Park, 2006; Fu et al., 2006) have emphasised the role of deep convective activities in bringing air from the upper troposphere to the lower stratosphere, another possible mechanism is that the divergence in the lower stratosphere associated with Tibetan anticyclone could be compensated by upward air mass transport. It is this slow process, rather than rapid convective activities, which carries ozone-poor tropospheric air into the lower stratosphere.

It is necessary here to further clarify whether chemical processes play a substantial role in causing the TCO low over the TP. Due to the CCM configuration, there is no chemistry in the CCM below 150 hPa, consequently, we cannot diagnose the role of the tropospheric chemistry in forcing the TCO low over the TP. However, tropospheric ozone makes only a small ($\sim$10%) contribution to the total column and our analysis of the TOR suggests that tropospheric processes only cause an ozone change of around 5 DU over the TP. Therefore, although the chemical processes in the troposphere could affect the TCO over the TP, the impact of modulation in tropospheric chemical processes on TCO is likely not significant. The question is whether the differences in ozone number density shown in Fig. 7a are also related to chemical processes in the UTLS. In fact, Fig. 10 indicates that larger ozone values over the TP are accompanied by the low methane values in the lower stratosphere and the same result can be noted in Fig. 7d. Since methane is a long-lived chemical species in the stratosphere it's variation is not subject to significant chemical modulation and changes in methane are mainly due to transport processes. The similar patterns exhibited in the ozone and methane distributions on the 450 K isentropic surface suggest that the TCO low over the TP is mainly caused by transport processes.

To provide further information on the chemical processes involved in the UTLS, Fig. 11 shows the CCM NO_x and CO fields on the 450 K isentropic surface. In our model CO is a marker of methane oxidation chemistry (recall that the model does not include tropospheric chemistry) while NO_x is an example of a family which leads to catalytic O_3 loss. In the lower stratosphere O_3 is destroyed chemically though its lifetime is relatively long (months–years). The most important lower stratosphere catalytic cycles are those involving HO_x and also halogen species. The NO_x loss cycle becomes more important towards the mid stratosphere. Before considering modelled chemical species we can point that because of the long chemical lifetime of ozone, it is not likely that local differences in chemical processes can contribute to the lower values of stratospheric ozone over the TP. Even if the local O_3 loss rate were to increase, the effects of this on the O_3 field will be transported to other longitudes by the zonal flow and lead to an overall widespread decrease in O_3. Although HO_x radicals are the most important for lower stratosphere ozone loss, due to the limitation of computer resources H, OH, and HO_2 fields were not output in our simulation. Nevertheless, the distribution of CO and NO_x on the 450 K surface

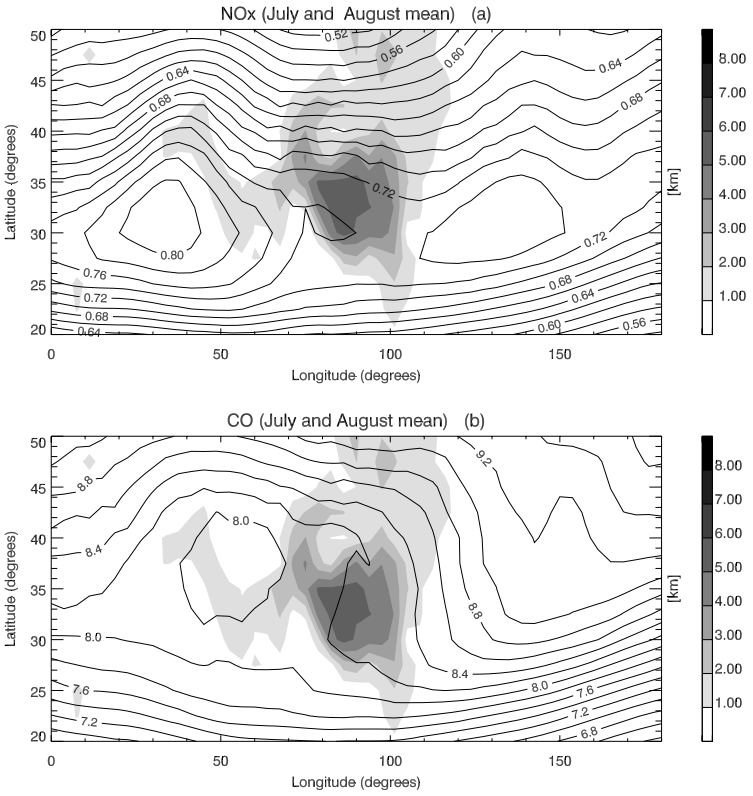

Fig. 11. Mean 20-yr CCM summer (July and August) latitude-longitude cross-sections on the 450 K isentropic surface for (a) NO_x (= NO + NO_2, contour interval 0.02 ppbv) and (b) CO (contour interval 0.2 ppbv). The filled contours represent topography (km).

can provide some information on possible local differences in chemical processes involved.

CO has a similar distribution on the 450 K surface as those of ozone and methane while NO_x distribution exhibits a different pattern. There are high values of NO_x in the WTP and the ETP compared to that over the TP implying the TP indeed exerts an influence on the chemical processes above it. However, the differences in NO_x between the TP and ETP are much smaller than those between the TP and the WTP. Recall that ozone differences between the TP and the ETP are larger than those between the TP and WTP. The result here supports the notion that chemical processes are not a major cause of the ozone differences between the TP and the ETP. In general, there is increased water vapour over the TP, and particularly over the ETP in the UTLS region as shown in Fig. 8b, one may expect that the NO_x-O_3-HO_x chemical processes are amplified by increased water vapour. However, this amplification does not look to be significant by the small variations of NO_x concentrations between the TP and the ETP. Further examination of Br_x field (not shown) indicates that Br_x distributions on the 450 K surface also has a similar pattern as those of O_3, CH_4 and CO. It is apparent that the tracer distributions on 450 K surface are dominated by the isentropic transport associated with the monsoon anticyclone circulation.

Finally, both the modelled ozone and SAGE II observations indicate that the differences are most pronounced below 20 km, indicating that the effect of the TP on ozone is not significant in the middle and upper stratosphere where the photochemical lifetime of ozone becomes short. The result here is consistent with the speculation of Moore and Semple (2005) who showed that the influence of the TP extends into the UTLS. Both the cold-point tropopause and thermal tropopause heights are not higher than 18 km, implying the differences in ozone and methane mixing ratios above 18 km should not be significantly affected by the convective activities but are related to the vertical shift of isentropic surfaces.

5. Summary

By using various observational and model data sets we have analysed the main causes of the low TCO over the TP and underlying mechanisms. The total column low over the TP is most significant from spring to summer due to the combined effects of high surface altitude (i.e. thin total air column) and thermal-dynamical forcing of the TP. The modelled ozone and methane fields as well as other chemical species on the 450 K isentropic surface suggest that chemical processes in the lower stratosphere play a only minor role in contributing to the relative TCO low.

The overall thinness of the air column contributes a significant part (up to ∼60%) to the low TCO over the TP. After that the thermal-dynamical forcing of the TP appears to makes the most important contribution to the low TCO relative to its surrounding area. This forcing of the TP first causes an increase in tropopause

height in summer and a decrease in winter. The variation of tropopause height over the TP is accompanied by large-scale uplift of isentropic surfaces in summer and descent in winter in the UTLS. This leads to upward and downward shifts of tracer vertical profiles of several kilometres and hence enhanced low TCO values over the TP.

It should be pointed out that the tropopause height over TP is impacted by both the dynamical lifting and the thermal maintenance by terrain-induced local circulations (including, e.g. forced ascent, thermal convection and air expansion) as well as monsoon systems. Summertime local deep circulations over the TP and the so-called Tibetan High act to maintain a higher tropopause due to the existence of anticyclone in the upper level. An important feature of the Tibetan High is the intense convective activities with divergence flow in the upper troposphere. The frequent convective activities and forced ascent of air over the TP may maintain a higher tropopause on one side and more divergence flow in the UTLS region on the other. However, further studies are needed to clarify the relative contribution of STE and horizontal divergence/slow vertical cross-tropopause transport to the TCO low above TP.

The thermal-dynamical forcing of TP can have an important effect on STE and hence, tracer vertical profiles over the TP. While the mechanism proposed by previous studies emphasise an important role to tracer transport by deep convective activity and the associated STE processes over the TP, it is also possible that the divergence in the lower stratosphere associated with the persistent monsoon anticyclone could be compensated by upward air mass transport. This slow transport process would carry ozone-poor tropospheric air into the lower stratosphere which would then mix in horizontally. It is also found that the monsoon anticyclone circulation causes an isentropic transport of trace gases from the higher latitudes towards over the TP, and hence modify tracer distributions in the lower stratosphere. The details of this issue merit further investigation.

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