Tibetan Plateau Impacts on Global Dust Transport in the Upper Troposphere

CHAO XU, YAOMING MA, KUN YANG, AND CHAO YOU

Key Laboratory of Tibetan Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, and CAS Center for Excellence in Tibetan Plateau Earth Science, and University of Chinese Academy of Sciences, Beijing, China

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ABSTRACT

Dust is a major component of atmospheric aerosol worldwide, greatly affecting regional and global climate. In this study dust aerosol optical depth (DAOD) and dust mass fluxes (DMF) were evaluated at different altitudes using measurements by the Cloud–Aerosol Lidar with Orthogonal Polarization (CALIOP) and ERA-Interim data from March through May (MAM) for the period 2007–16. Significantly higher upper-tropospheric (above ~8 km) dust loads and DMF downstream of the Tibetan Plateau (TP) relative to those over other major dust sources of the Northern Hemisphere were found during spring. A DMF magnitude of 10^{10} g integrated across a 2° -latitude segment during spring was estimated downstream of the TP in the upper troposphere. A dust belt can be clearly seen at altitudes higher than 6 km over the downwind direction of the TP at latitudes of around 30° – 40° N, crossing the Pacific Ocean and extending to North America during spring. A pathway for transporting dust aerosols into the upper troposphere is proposed, as follows. Dust is uplifted to the midtroposphere over the source regions; then, frequent, deep, dry convection prevailing over the TP during spring can cause convective overshooting that uplifts the dust aerosols to the upper troposphere. The TP thus acts as a channel for transporting dust from the lower atmosphere to the upper troposphere, enabling the long-range zonal transport of dust around the Northern Hemisphere.

1. Introduction

Dust plays an essential role in the global climate system, biogeochemical cycles, human society, and ecosystems (e.g., Choobari et al. 2014; Huang et al. 2014, 2006; Kaufman et al. 2002; Li et al. 2011). Dust aerosols impact the radiation budget by scattering and absorbing solar radiation (Huang et al. 2014); they can also interact with clouds, alter cloud microphysical processes, and further impact the hydrological cycle (Huang et al. 2014; Li et al. 2016). Global-scale long-range transport of dust provides nutrients for the ocean, which are especially important for marine ecosystems (Yu et al. 2015a). Moreover, dust can strongly interact with the Asian monsoon system. It has been reported that an increase in dust from the Arabian Sea, western Asia, and the Arabian Peninsula could enhance the intensity of the Indian

summer monsoon and associated rainfall over central India on time scales of days to weeks (Jin et al. 2016; Lau 2014; Lau and Kim 2016; Li et al. 2016; Vinoj et al. 2014).

A dust belt, extending from the west coast of North Africa, passing across the Arabian Peninsula and central and southern Asia, to northern China, has been identified in the Northern Hemisphere (Ginoux et al. 2012; Prospero et al. 2002). Much attention has been dedicated to understanding the global-scale long-range transport of dust, including its transatlantic and transpacific transport (e.g., Ben-Ami et al. 2012; Bourgeois et al. 2015; Huang et al. 2010; Yu et al. 2015a). Dust originating in North Africa, the Arabian Peninsula, and southwestern Asia is observed to peak in summer, while that from central and East Asia peaks in spring (Alizadeh-Choobari et al. 2014; D. Liu et al. 2008). African dust is transported across the Atlantic throughout the year, with transport pathways mainly occurring in the free troposphere and showing clear seasonal variations (D. Liu et al. 2008). Asian dust can be transported thousands of kilometers to remote, pristine regions, including Greenland (Bory et al. 2003) and the Tibetan Plateau (TP) (Bucci et al. 2014; Huang et al. 2007; Z. Liu et al. 2008). Asian dust also contributes most of the dust load in the troposphere over the

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Corresponding author: Chao Xu, xuchao@itpcas.ac.cn; Yaoming Ma, yyma@itpcas.ac.cn

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midlatitude regions from East Asia to western North America during spring (Zhao et al. 2006). It has been observed that dust originating from the Taklamakan Desert can be transported through one full global circuit within 13 days (Uno et al. 2009).

The residence time of dust aerosols in the long-range transport system is largely related to their vertical location (Bourgeois et al. 2015). Therefore, the characteristics of the vertical distribution of dust aerosols are crucial to comprehensively understanding the long-range transport of dust. Once dust aerosols enter the upper troposphere (above ~ 8 km), they can be transported around Earth in a zonal belt in a week or two via upper-tropospheric westerly jets (Huang et al. 2008). Although the global transport of dust has been observed or simulated, the primary mechanism of dust transport in the upper troposphere is still poorly understood. It has also been reported that the direct incursion of tropospheric air even into the lower stratosphere as a result of fast convective overshooting updrafts occurs on the TP (Fu et al. 2006). Therefore, it is worth assessing the role played by the TP in the transport of dust into the upper troposphere in the Northern Hemisphere.

The TP, often referred to as the third pole, is located in central eastern Eurasia and has the world's highest average elevation (about 4000 m), with some surface features even reaching the midtroposphere (Yao et al. 2012). Because of its topographic characteristics, the TP surface absorbs high quantities of solar radiation with corresponding impacts on surface heat or water fluxes (Ma et al. 2014a,b). The strong thermodynamic activity of the TP directly affects the climate system in the Northern Hemisphere (Li and Zhang 2012; Wu et al. 2015, 2007). Figure 1 illustrates the global 10-yr averages of column dust aerosol optical depth (DAOD). Several dust sources are distributed around the windward side of the TP, including North Africa, the Arabian Peninsula, the Iranian Plateau, and the Tarim basin, and the TP usually acts as a receptor of dust. Dust has been found to be the most prominent aerosol type on the TP, especially during spring (Kuhlmann and Quaas 2010; Xu et al. 2015). In this study DAOD and dust mass fluxes (DMF) were evaluated at different altitudes using data from the Cloud–Aerosol Lidar with Orthogonal Polarization (CALIOP) and ERA-Interim from March through May (MAM) for the period 2007–16. This study aims to reveal the impact of TP on the long-range transport of dust in the upper troposphere in the Northern Hemisphere.

2. Data and methodology

a. CALIPSO

Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) provides new insights into

clouds and atmospheric aerosols (Winker et al. 2007, 2010). CALIOP is an instrument on board the *CALIPSO* satellite. The *CALIPSO* satellite was launched into a sun-synchronous orbit on 28 April 2006, with a 16-day repeating cycle. The lidar level-3 aerosol product is a quality-screened aggregation of level-2 aerosol profile data. A series of filters has been designed to eliminate samples and layers that were detected or classified with very low confidence or that have untrustworthy extinction retrievals (Winker et al. 2013). The CALIOP instrument is more sensitive during nighttime than daytime. Solar background illumination decreases the signal-to-noise ratio (Winker et al. 2010). The classification algorithms use integrated attenuated backscatter measurements, volume depolarization ratio measurements, surface type, and layer altitude to determine aerosol type (Omar et al. 2009). Previous studies have demonstrated the reliability of CALIOP measurements by comparison with other observation techniques. The Aerosol Robotic Network (AERONET) is a federated international global network of surface aerosol measurement stations, and the strong performance of *CALIPSO* has been demonstrated by comparison with AERONET data (Schuster et al. 2012). Mielonen et al. (2009) verified that the classifications of dust by CALIOP were reliable in comparison with AERONET data, achieving good agreement for the dust type (91% of the cases). Burton et al. (2013) also confirmed the reliability of dust classifications when comparing with airborne high spectral resolution lidar, obtaining good agreement for dust (80% of the cases).

The *CALIPSO*, version 3.00, level-3 nighttime aerosol profile data during MAM for the period 2007–16 (available online at https://eosweb.larc.nasa.gov/project/calipso/calipso_table) were used in this study, and the primary variable was the dust extinction at 532 nm (km^{-1}). The spatial resolution of level-3 data is $5^\circ \times 2^\circ$ (longitude $\times$ latitude), and the vertical resolution is 60 m from -0.5 to 12 km above mean sea level (MSL; all altitudes hereafter also refer to MSL) in the troposphere.

We calculated DAOD at different altitudes from the dust extinction at 532 nm as follows:

$$\text{DAOD} = \int_{z_1}^{z_2} \text{DE}(z) dz, \quad (1)$$

where $\text{DE}(z)$ is the dust extinction at 532 nm (km^{-1}) at altitude z , and z_1 and z_2 are the altitude range of DAOD, in 12 layers at 1-km intervals from the surface to 12 km.

To assess the long-range transport of dust, we focus on zonal transport at four height–longitude cross sections along 16° , 24° , 36° , and 38°N , where 16°N primarily crosses North Africa; 24°N crosses North Africa, the

FIG. 1. The global 10-yr average of column DAOD during MAM 2007–16. NA indicates North Africa, AP indicates Arabian Peninsula, IrP indicates Iranian Plateau, InP indicates Indian subcontinent, TB indicates Tarim basin, TP indicates the Tibetan Plateau, PO indicates the Pacific Ocean, and AO indicates the Atlantic Ocean (these abbreviations are used repeatedly in the figures).

Arabian Peninsula, and Indian subcontinent; and 36° and 38°N mainly cross the TP. Dust extinction at 532 nm was averaged along the four height–longitude cross sections over 2°-latitude segments during MAM from 2007 to 2016. To better describe where dust is concentrated, the altitudes with dust extinctions higher than 0.001 and 0.0001 km^{−1} were also depicted. To eliminate abrupt changes, the altitudes of dust layers meeting the abovementioned two thresholds were calculated as moving averages of three points.

The DMF were estimated using the method of Yu et al. (2015a) as follows. First, we converted dust extinction at 532 nm (km^{−1}) to dust mass concentration m (g m^{−3}) with the assumed dust mass extinction efficiency (MEE) of 0.37 m² g^{−1}, following Kaufman et al. (2005) and Yu et al. (2015a). We calculated the dust mass flux rate (FR; g s^{−1}) in the zonal (east–west) direction in each 5° × 2° grid cell as follows:

$$\text{FR} = \int_{z_1}^{z_2} m(z)U(z)L dz, \quad (2)$$

where $m(z)$ is the mass concentration (g m^{−3}) at altitude z , z_1 and z_2 are the altitude range of the flux rate, L is the length (m) of a 2°-latitude segment along a line of longitude, and U is the zonal wind speed (m s^{−1}) retrieved from ERA-Interim data (Berrisford et al. 2011; Dee et al. 2011). DMF were retrieved in 12 layers at 1-km intervals from the surface to 12 km. Monthly DMF were calculated by multiplying the monthly average flux rate and the duration time (s) in that month. DMF in spring was calculated by summing the DMF during MAM in each year, and then the multiyear average of DMF in

spring was calculated by averaging spring DMF from 2007 to 2016. Eastward transport is defined as a positive flux and westward transport as a negative flux. The estimated uncertainties in DMF are associated with CALIOP extinction, vertical profile shape, dust mass extinction efficiency, and possible changes in the dust size distribution during transport. The cumulative uncertainty of all these error sources has been estimated to be ±45%–70%, while uncertainties related to the diurnal variation of dust and the missing below-cloud dust cannot be quantified because of a lack of reliable observations (Yu et al. 2015a).

b. ECMWF data

The meridional circulation at 90°E and the zonal circulations at 24° and 36°N were examined using monthly mean ERA-Interim data during MAM for the period 2007–16, available from the European Centre for Medium-Range Weather Forecasts (ECMWF) (Berrisford et al. 2011; Dee et al. 2011). Daily vertical velocity was retrieved from daily ERA-Interim data, produced by ECMWF (Berrisford et al. 2011; Dee et al. 2011). The spatial resolution of monthly and daily ERA-Interim data is 0.5° × 0.5° (available online at <http://apps.ecmwf.int/datasets/>). The occurrence frequencies of updrafts were evaluated using the daily instantaneous vertical velocity (Pa s^{−1}) at four different times each day (0000, 0600, 1200, and 1800 UTC) during MAM 2007–16. The occurrence frequencies of updrafts $f_{\text{occurrence}}$ at each time were defined as the ratio of the number of days with upward vertical velocity N_{updrafts} [vertical velocity < 0 at all pressure levels from 500 (about 5.5 km) to 300 hPa (about 9.5 km) at 50-hPa intervals] to the total number of days N_{total} as follows:

$$f_{\text{occurrence}} = \frac{N_{\text{updrafts}}}{N_{\text{total}}}. \quad (3)$$

3. Results

a. DAOD at different altitudes

The global 10-yr averages of DAOD at different altitudes during the major dust active season (MAM), derived from CALIOP measurements during 2007–16, are displayed in Fig. 2. DAOD between 0 and 1 km, between 3 and 4 km, between 6 and 7 km, and between 9 and 10 km was selected as being representative of the variations in dust at different altitudes; DAOD at other altitudes from the surface to 12 km at 1-km intervals is shown in Fig. S1 of the supplemental material. Arid and semiarid regions, including the Sahara and Sahel, the Arabian Peninsula, northern India, the Tarim basin, and

FIG. 2. The global 10-yr averages of DAOD between 0 and 1 km, between 3 and 4 km, between 6 and 7 km, and between 9 and 10 km during MAM 2007–16.

the Gobi Desert, are the major dust sources, comprising the dust belt in the Northern Hemisphere (Ginoux et al. 2012; Huang et al. 2016; Prospero et al. 2002). However, significant differences are revealed between the lower troposphere (below ~ 3 km) and the upper troposphere (above ~ 8 km).

Globally, DAOD enhancements occur mainly in the near-surface layers (below ~ 3 km) around the dust sources. DAOD over the major dust sources can exceed 0.1 between 0 and 1 km. Between 3 and 4 km, the Tarim basin has a greater dust load than North Africa and the Arabian Peninsula. DAOD decreases notably with increasing altitude over the source regions, typically decaying by one order of magnitude from between 3 and 4 km to between 6 and 7 km. DAOD enhancements between 9 and 10 km appear to be unrelated to the source regions (Fig. 2).

It is noticeable that DAOD between 4 and 6 km over the TP is lower than that over the strong source regions listed above (Fig. S1). Conversely, DAOD between 9 and 10 km reached higher values over the northern TP and downstream of the TP than those over the major dust source regions (Fig. 2). Similar magnitudes of DAOD were not achieved in the upper troposphere over upstream regions of the TP, indicating that dust can be more readily lofted to the

upper troposphere over the TP than over other dust source regions. A dust belt can be observed extending from the downstream area of the TP across to North America in the upper troposphere at latitudes of around 30° – 40° N.

b. Vertical structure of dust extinction

Figure 3 illustrates the 10-yr averages of the vertical distribution of dust extinction along some typical height–longitude cross sections during MAM, to depict the zonal transport of dust. Although the Sahara Desert is generally the strongest dust source globally (e.g., D. Liu et al. 2008; Prospero et al. 2002), more intense dust layers were found over the TP and downstream of the TP relative to the Sahara Desert.

Strong dust extinction, exceeding 0.1 km^{-1} , can be observed over the source regions. The maximum height of the dust layer (dust extinction $> 0.001 \text{ km}^{-1}$) is around 6 km along 16° and 24° N. Strong dust peaks are mainly located over North Africa, the Arabian Peninsula, and the Indian subcontinent along both cross sections. During transport from North Africa to the Atlantic, dust plumes consistently descend and the dust extinction also decreases significantly with altitude (Liu et al. 2012). Therefore, strong dust extinction is found only over the source regions, and dust

FIG. 3. The 10-yr averages of dust extinction for height–longitude cross sections at 16°, 24°, 36°, and 38°N during MAM 2007–16. The height–longitude cross sections are over 2°-latitude segments, with the indicated latitude at the center of the cross section. The black solid line indicates the heights where dust extinction exceeds 0.0001 km^{-1} , and the blue solid line indicates the heights where dust extinction exceeds 0.001 km^{-1} . The respective location maps are shown above each panel. The red lines indicate the location of the cross sections.

plumes descend quickly during their subsequent transport.

The height–longitude cross sections along 36° and 38°N cross the TP. Dust layers with dust extinctions higher than 0.001 and 0.0001 km^{-1} display consistent

patterns. The heights of the dust layers (dust extinction $>0.001 \text{ km}^{-1}$) begin to increase after crossing the TP in the 36° and 38°N transects. The TP appears to alter the transport route of dust from low levels in the west to higher levels in the east, in both transects. It is

FIG. 4. The global 10-yr averages of DMF (g) integrated across a 2° -latitude segment between 0 and 1 km, between 3 and 4 km, between 6 and 7 km, and between 9 and 10 km during MAM 2007–16. Positive DMF indicate eastward transport and negative DMF indicate westward transport.

important to note that the altitude of the dust layers peaks on the leeward side of the TP in these two cross sections, perhaps because the dust is forced over the TP by frequent convective updrafts and retains an upward velocity after passing the TP. Dust may continue rising for some time after the TP before descending during its subsequent eastward transport. Therefore, the highest dust layer forms to the east of the TP. The top layer of dust (dust extinction $>0.0001 \text{ km}^{-1}$) is higher than 8 km over eastern China and the western Pacific (120°E – 180°), sometimes even reaching 10 km (Fig. 3). Dust plumes descend gradually over the Pacific and North America with increasing transport distance (Yi et al. 2014). Previous studies have indicated that dust from the Middle East and North Africa could be transported to the Pacific by the westerlies (Hsu et al. 2012; Tanaka et al. 2005). North African dust is primarily transported toward South America in the lower atmosphere (Glaser et al. 2015; Yang et al. 2012), while upper tropospheric westerly jets may allow the long-range transport of North African dust aerosols (Hsu et al. 2012). Both the lowest dust extinction

and the lowest dust layers appear over the Atlantic, and dust plumes even descend to the lower troposphere (sometimes lower than 2 km), indicating that dust transport leads to large quantities of dust deposition into the ocean.

Generally, dust can be elevated to the midtroposphere (below 7 km) over the source regions. However, dust layers (dust extinction $>0.001 \text{ km}^{-1}$) over the downstream areas of the TP can increase in height, even exceeding 8 km. The top layer of dust reaches as high as 10 km. This leads us to ask what process causes the highest dust layers to occur to the lee of the TP, rather than over the source regions, in the Northern Hemisphere.

c. DMF at different altitudes

To better understand global dust transport, it makes sense to study the DMF (e.g., Huang et al. 2015; Kaufman et al. 2005; Yu et al. 2015a,b). Figure 4 illustrates the representative global 10-yr averages of DMF integrated across a 2° -latitude segment between 0 and

1 km, between 3 and 4 km, between 6 and 7 km, and between 9 and 10 km during MAM, while the DMF at other altitudes from the surface to 12 km at 1-km intervals is shown in Fig. S2 of the supplemental material. Similar to the results of DAOD (Fig. 2), there are clear differences in DMF between the lower and upper atmosphere. Regions of strong DMF appear in the lower troposphere (below ~ 3 km) around dust source regions, whereas relatively high regions of DMF in the upper troposphere (above ~ 8 km) appear unrelated to major dust sources (Fig. 4).

DMF decrease greatly with increasing altitude above 4 km over the major dust source regions. Importantly, the dust belt at latitudes of around 30° – 40° N begins to emerge between 6 and 7 km. The magnitude of DMF downstream of the TP is around 10^{11} g integrated across a 2° -latitude segment between 6 and 7 km, while the magnitude between 9 and 10 km decreases to 10^{10} g integrated across a 2° -latitude segment. Higher DMF can be found downstream of the TP relative to other major dust sources in the upper troposphere.

It is also worth noting that DMF show westward transport in the lower troposphere (below ~ 3 km) over the Tarim basin, as result of local circulations, switching to eastward transport at altitudes higher than 3 km (Figs. 4 and S2). A strong anticyclonic wind anomaly at 500 hPa and enhanced easterly wind at 850 hPa over the Tarim basin have been shown to promote dust entrainment, vertical lofting, and horizontal transport (Ge et al. 2014). North African dust also presents westward transport between 0 and 1 km, switching to eastward transport between 3 and 4 km. To the south of 15° N over Africa, DMF almost always showed negative values at the altitudes below 6 km (Figs. 4 and S2), consistent with the results of Yu et al. (2015a). Moreover, the magnitude of DMF in both the Tarim basin and North Africa exceeded 10^{12} g integrated across a 2° -latitude segment in the lower troposphere.

d. Possible explanation of the dust belt in the upper troposphere

Atmospheric circulation greatly impacts the dust transport. Figures 5 and 6 show the 10-yr averages of the spring meridional and zonal circulation, respectively. The latitude–pressure (altitude) cross section along 90° E and the longitude–pressure (altitude) cross section along 24° and 36° N were selected to analyze the vertical atmospheric circulation over the TP. During spring the northern and southern air flows intersect at about 32° – 33° N over the TP in the cross section along 90° E (Fig. 5). Updrafts can be observed over the TP from the surface to as high as 100 hPa, and the seasonal means of vertical velocity above the TP

FIG. 5. The 10-yr averages of meridional circulation (arrows) and vertical velocity (Pa s^{-1} ; shading) at 90° E during MAM 2007–16. The shading shows the topography at 90° E (black) and indicates updrafts (blue) and downdrafts (red).

can reach from -0.04 to approximately -0.10 Pa s^{-1} below 300 hPa (Fig. 5 and Fig. 6, right). The vertical velocity seems to be higher above the peripheral slopes than above the main body of the TP, possibly as a result of slope heating (Wu et al. 2012). Meanwhile, dust that accumulates on the slopes of the TP can absorb more solar radiation and then enhance the slope heating, according to the elevated heat pump hypothesis (Lau et al. 2006, 2008). The dynamic effects of the TP, for instance the climbing flows, are also crucial for the formation of updrafts. It has been reported that the large-scale orography of the TP affects the Asian climate through thermal forcing in spring and summer, and through mechanical forcing in winter (Wu et al. 2015). Frequent updrafts over the TP enable dust transportation to the upper troposphere.

The vertical atmospheric circulations across North Africa (0° – 30° E) can be seen in the cross section along 24° N (Fig. 6, left). The Sahara Desert is generally considered as the strongest and largest dust source globally (e.g., D. Liu et al. 2008; Prospero et al. 2002). However, updrafts mainly occur below 450 hPa over the western part of North Africa (0° – 15° E) (Fig. 6, left). Therefore, dust can be transported only to the midtroposphere over North Africa.

The diurnal frequency of updrafts from the 500- to 300-hPa levels was evaluated using daily ERA-Interim data during MAM 2007–16 (Fig. 7), and the method for calculating the frequency of updrafts is shown in section 2b. North Africa, the Arabian Peninsula, the Iranian Plateau, and the Tarim basin are commonly identified as the major dust sources in the Northern Hemisphere.

FIG. 6. The 10-yr averages of zonal circulation (arrows) and vertical velocity (Pa s^{-1} ; shading) at 24° and 36°N during MAM 2007–16. Respective location maps are shown above each panel. The shading shows the topography at 24° and 36°N (black) and indicates updrafts (blue) and downdrafts (red).

Because of large changes in time zone, the same coordinated universal time (UTC) indicates different local times (LT) in these regions (local time for North Africa is approximately UTC, the Arabian Peninsula is approximately UTC + 3 h, the Iranian Plateau is approximately UTC + 4 h, and the Tarim basin and TP are approximately UTC + 6 h). The frequency of updrafts over most of the Sahara Desert is below 30% throughout the day, and it occasionally reaches 45% at 1200 LT in small regions of the western Sahara. The frequency of updrafts increases slightly over the Arabian Peninsula and Iranian Plateau, with the highest frequency of updrafts at 1500–1600 LT, occasionally reaching 45% over part of the Arabian Peninsula and Iranian Plateau. The frequency of updrafts is nearly below 30% throughout the day over the Tarim basin. The frequency of updrafts is found to be high over the TP relative to the above-mentioned source regions, and sometimes even comparable to those in the tropics. The updrafts show a clear diurnal pattern over the TP, with the greatest frequency of updrafts during 1200–1800 LT. Continuous updrafts occur from 500 to 300 hPa over the main body of the TP in more than 45% of cases at 1200 LT. Moreover, the frequency of updrafts at 1800 LT is around 60% over the main body of the TP, sometimes even reaching 66% of cases. Regardless of the dust origin, frequent updrafts over the TP can carry dust to the upper troposphere, from where it is then transported globally via upper-tropospheric westerly jets.

A relay-transport process is proposed to explain the effect of the TP in the formation of the dust belt over the Northern Hemisphere (Fig. 8). Dust can usually be elevated to the midtroposphere (less than 7 km) over the dust source regions (Figs. 2 and 4). During the spring the land surface of the TP is a strong source of sensible heat to warm the surface air (Ma et al. 2011). This causes cyclonic conditions in the lower layer and draws dust-laden air from neighboring regions onto the TP. Meanwhile, strong sensible heat in this season often triggers deep, dry convection over the TP (Yang et al. 2004). Under these conditions the TP as a whole can uplift dust from the midtroposphere to the upper troposphere (Fig. 8). An anticyclone dominates the upper troposphere over the TP, and TP warming can strengthen this anticyclone in the upper levels. The TP warming not only promotes local convection but also greatly affects the subsequent dust transport through the Rossby wave train. It has been found that the Rossby wave train is excited by the anomalous thermal heating over the TP (Wang et al. 2008). In brief, the circulation anomaly is generated in response to the forcing by the TP warming. The vorticity wave train propagates downstream and eastward along the upper-level westerly jet stream. The cyclonic circulation in the upper troposphere develops over East Asia as a result of the TP warming, and the forced downdrafts facilitate dust deposition back to Earth's surface. Propagation of the wave train enhances the anticyclonic circulation in the upper troposphere to

FIG. 7. Occurrence frequencies of updrafts from the 500- to 300-hPa levels at 0000, 0600, 1200, and 1800 UTC during MAM 2007–16. The method using daily ERA-Interim data for calculating occurrence frequencies of updrafts is detailed in [section 2b](#).

the east of Japan. The forced airflows to the east of Japan may contribute to the dust uplift, aiding the transport of dust across the North Pacific and even globally. Wet deposition predominates over the transpacific transport pathway, while dry deposition is dominant near the source areas ([Zhao et al. 2006](#)). Of course, several dust sources are located in the windward regions of the TP and are a further prerequisite for the upper-tropospheric dust transport triggered by the TP. Therefore, the stable dust belt in the upper troposphere downstream of the TP is a reliably persistent feature, because of the abovementioned processes.

4. Conclusions

This study calculated the global 10-yr averages of DAOD and DMF during MAM at different altitudes using CALIOP data for the period 2007–16. Great differences between the lower troposphere (below ~ 3 km) and upper troposphere (above ~ 8 km) were revealed in both DAOD and DMF. Higher DAOD and DMF in the lower troposphere were largely associated with major dust sources in the Northern Hemisphere, whereas

higher DAOD and DMF in the upper troposphere were observed downstream of the TP. Furthermore, the highest dust layers were typically seen over and to the lee of the TP along the zonal transects. DAOD can exceed 0.1 over the major dust sources in the lower troposphere, and DAOD downstream of the TP in the upper troposphere was around 0.0001. The magnitude of DMF during spring over the major dust sources was around 10^{12} g integrated across a 2° -latitude segment in the lower troposphere, reaching 10^{10} g integrated across a 2° -latitude segment downstream of the TP in the

FIG. 8. Schematic diagram illustrating the formation of the dust belt over the Northern Hemisphere. The letters A and C denote the anticyclonic and cyclonic circulation centers, respectively.

upper troposphere. A dust belt was evident at latitudes of around 30°–40°N and altitudes higher than 6 km.

Atmospheric circulation patterns impact the dust transport. Updrafts can be clearly observed over the TP, even at 100 hPa. The seasonal means of vertical velocity above the TP can reach from -0.04 to approximately -0.10 Pa s^{-1} below 300 hPa. A pathway for transporting dust aerosols into the upper troposphere is proposed, as follows. Dust is uplifted to the midtroposphere over the source regions; then, frequent, deep, dry convection prevailing over the TP during spring can uplift the dust aerosols to the upper troposphere. This study provides a preliminary view of dust characteristics and proposes a pathway for transporting dust aerosols into the upper troposphere. The findings are of great importance for understanding the formation of the dust belt over the Northern Hemisphere, and they also provide a theoretical basis for dust transport in global-scale climate modeling.

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