The impact of atmospheric stability and wind shear on vertical cloud overlap over the Tibetan Plateau

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Abstract

The accurate representation of cloud vertical overlap in atmospheric models is important for simulating the total cloud cover and the radiative energy budget in these models. However, this subject has received little attention due to the limited observation, especially over the Tibetan Plateau (TP), where has experienced a rapid climate warming over the past three decades. In this study, 4 years (2007–2010) of data from the CloudSat cloud product and collocated ERA-Interim reanalysis product are analyzed to examine the cloud overlaps over the TP region, and evaluate the effect

of atmospheric dynamic and thermodynamic environment on these cloud overlaps. The overlap parameter α and decorrelation length scale *L*, which are widely used to characterize the transition from the maximum to random overlap assumption with increasing layer separations, are calculated and discussed. It is confirmed that continuous cloud layers tend to have a maximum overlap at a small separation but gradually become randomly overlapped with increasing cloud layer separations. It is found that for the continuous cloud layers, the overlap parameter α is sensitive to the unique thermo-dynamic and dynamic environment over the TP, i.e., the unstable atmospheric stratification and corresponding weak wind shear, which leads to maximum overlap (that is, greater α values). This finding agrees well with the previous studies. We parameterize the decorrelation length scale *L* as a function of the wind shear and atmospheric stability based on a multiple linear regression. Compared with previous parameterizations, this new scheme improves the simulation of cloud cover over TP when the separations between cloud layers are larger than 1km. This study indicates that effects of both wind shear and atmospheric stability on cloud overlap should be taken into account in the parameterization of overlap parameter α to improve the simulation of total cloud cover in atmospheric models.

1. Introduction

Clouds strongly modulate the Earth's radiative energy budget, via changes in their macrophysical (e.g., cloud cover, height, and thickness) and microphysical properties (e.g., cloud water contents, phase and droplet and crystal size) (Rossow and Lacis, 1990; Hartmann et al., 1992; Fu and Liou, 1993; Fu et al., 2002; Stephens, 2005; Kawamoto and Suzuki, 2012; Yan et al., 2012; Wang et al., 2010). However, our incomplete understanding of their underlying physical processes makes the representation of clouds in climate models still unreliable, which keeps clouds as the largest uncertainty when estimating and interpreting changes in the Earth's energy budget (Boucher et al., 2013).

The Tibetan Plateau (TP), which is also known as the "roof of the world" or the "world water tower", plays a significant role in determining global atmospheric circulations, in addition to its strong influence over Asia via its thermal-dynamic and dynamic forcings (Yanai et al., 1992; Ye and Wu, 1998; Duan and Wu, 2005; Xu et al., 2008; Wu et al., 2015). Over the past three decades (Kang et al., 2010), the TP has experienced distinct climate changes including the changes in atmospheric circulations and hydrological cycles (Yang et al., 2014). Many studies have showed that significant warming occurs in the TP region during the last decades and it will continue in the future (e.g., Duan et al., 2006; Wang et al., 2008). The rapid warming has caused glacier retreat and expansion of glacier-fed lakes (Zhu et al., 2010), permafrost degradation (Cheng and Wu, 2007), heating source has became weakened (Yang et al., 2011). The warming and weakened heating source also affects the summer precipitation downstream (Duan et al., 2013). In a review paper, Kang et al. (2010) summarized that the changes of cloud cover based on observations also is one of dominant factors causing the rapid warming over TP region in addition to increased greenhouse gas emission. Many studies indeed have linked the rapid warming over TP to changes in the cloud cover over the TP region (e.g., Chen and Liu, 2005; Duan and Wu, 2006; Li et al., 2006; Yang et al., 2012; You et al., 2014; Wu et al., 2014). A recent study has indicated that increased nocturnal cloud cover over the northern TP could increase the nighttime temperature by enhancing downward surface infrared

radiation, while decreased daytime cloud over the southern TP has contributed to the increase of surface air temperature during daytime by enhancing downward surface solar radiation (Duan and Xiao, 2015). Because of the importance of clouds in climate change, it is critically important to reliably represent the cloud cover and its relation to large-scale thermodynamic and dynamic conditions in the climate models in order to better predict the climate changes over TP.

However, our understanding about the role of cloud cover on the radiation balance and water cycle over the TP region remains poor because of the limited availability of regional cloud observations and our incomplete representation of the cloud physical processes in climate models. One of the remaining challenges involves how to reasonably represent the characteristics of the vertical overlapping of cloud layers in these models. Cloud overlap means that two or more cloud layers are simultaneously present over the same location but at different levels in the atmosphere. It is usually defined in terms of three basic overlap assumptions: maximum, random and minimum. If the cloud cover in two model layers is given by C_i and C_j, respectively, total cloud cover of any two cloud layers from maximum assumption is $C_{i,j}^{\max} = \max\{C_i, C_j\}$, the random and minimum assumptions give the total cloud cover as $C_{i,j}^{ran} = C_i + C_j - C_i \times C_j$ and $C_{i,j}^{\min} = \min\{C_i + C_j, 1\}$,respectively. Thus, the maximum assumption minimizes the total cloud cover, while minimum assumption produces minimally overlap between cloud layers and results in maximum total cloud cover (Weger et al., 1992). The total cloud cover predicted by the random assumption will fall somewhere between maximum and minimum assumption (Geleyn and Hollingsworth, 1979). For example, if the cloud covers in two model layers are 50%, then the maximum overlap will result in a total cloud cover of 50%, and a minimum overlap will result in an overcast condition (a complete cloud cover, i.e., 100%). These different overlap assumptions result in obvious different total cloud covers and will significantly affect the calculated radiative budgets and heating/cooling rate profiles (Morcrette and Fouquart, 1986; Barker et al., 1999; Barker and Fu, 2000; Chen et al., 2000; Pincus et al., 2005; Zhang et al., 2013a; 2013b; 2016; Jing et al.,

2016). Previously studies based on general circulation model (GCM) simulations indicated that the bias in the global mean radiation fluxes at the top of the atmosphere and at the surface can reach 20-40 W m^{-2} due to the different overlap treatments (Morcrette and Jakob, 2000; Jing et al., 2009; Zhang and Jing, 2010).

Hogan and Illingworth (2000) revisited the cloud overlap assumptions and proposed a simpler and more useful expression for the degree of cloud layer overlap (exponential random overlap assumption) by using the ground-based radar measurement. In the expression, the observed cloud fraction of two cloud layers can be expressed as the linear combination of the maximum and random overlap by using a weighting factor, termed as cloud overlap parameter α:

$$\alpha = \frac{C_{i,j}^{obs} - C_{i,j}^{ran}}{C_{i,j}^{\max} - C_{i,j}^{ran}} \quad (1)$$

The overlap parameter ranges from 0 (random) to 1 (maximum) when the observed total cloud cover falls between the values using the maximum and random overlap assumptions. The α will be negative when the degree of cloud overlap is lower than that predicted by the random overlap assumption. In the study of Hogan and Illingworth (2000), they also fitted the reduction in α with layer separation distance D as an inverse exponential function of the decorrelation length scale L: $\alpha = e^{-D/L}$. Until now, many efforts that have been made to characterize the overlap parameter α and decorrelation length L using ground-based radar observations (e.g. Mace and Benson-Troth, 2002; Willén et al., 2005; Naud et al., 2008; Oreopoulos and Norris, 2011). For example, Oreopoulos and Norris (2011) derived L based on radar measurement taken over the US Southern Great Plains (SGP). Their results indicated that L ranges from 2 to 4.5 km across different seasons and smaller spatial scales correspond with smaller L values. Although radar can provide reliable cloud vertical structure profiles continuously, these observations are for a given location (Ge et al., 2017; 2018), and the radar sites are very sparsely distributed over the world, especially over the TP region where long-term Radar observations are nonexistent. Passive sensors and traditional surface weather reports fail to detect vertical cloud

structures, and only provide limited information about the cloud overlap (Chang and Li, 2005a, b; Huang, 2006; Huang et al., 2005, 2006a). The millimeter-wavelength cloud profiling radar (CPR) launched on CloudSat (Stephens et al., 2002) and the cloud-aerosol lidar with orthogonal polarization (CALIOP) (Winker et al., 2007) launched on CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation) provide an unprecedented opportunity to investigate vertical cloud overlaps on a global scale and to improve model representation of *L* (Barker et al., 2008; Kato et al., 2010; Mace et al., 2009; Li et al., 2011; 2015; Shonk et al 2010; Shonk et al., 2014; Tompkins and Di Giuseppe, 2015; Di Giuseppe and Tompkins, 2015). Based on two months of cloud mask profile information from the CloudSat and CALIPSO satellites, Barker (2008) quantified the properties of cloud overlap on a global scale and found a wide range of *L* values, with a median value of 2 km. In other studies, *L* was usually a function of latitude or total cloud cover (Shonk et al., 2010; 2014; Yoo et al., 2014). Recently, Di Giuseppe and Tompkins (2015) further evaluated the impact of wind shear on the global-scale cloud overlap and identified an empirical relationship between the decorrelation length *L* and wind shear for use in models by using 6 months of CloudSat-CALIPSO data.

However, the related question of the cloud overlapping over the TP region has received little attention. It is still an open question on how the unique thermo-dynamic and dynamic environment over the TP region affects cloud overlap there. This study investigates the cloud overlap and its relation to the atmospheric states and large-scale atmospheric dynamics over the TP region by combining the cloud cover profile information from the 2B-GEOPROF-LIDAR dataset (Mace et al., 2009; Mace and Zhang, 2014) and the meteorological fields from the ERA-Interim reanalysis datasets (Dee et al., 2011). This paper is organized as follows. The datasets and methods used in this study are briefly described in Section 2. Section 3 outlines the monthly and zonal variations of the cloud overlap parameters over the TP region. The impacts of the atmospheric state and large-scale atmospheric dynamics on cloud overlap are presented in Sections 4. The conclusions and discussion are given in Section 5.

2. Datasets and methods

4 years (2007–2010) of data from the CloudSat 2B-GEOPROF-LIDAR, ECMWF-AUX and the daily 6-hour ERA-Interim reanalysis were used to analyze the impacts of atmospheric states and dynamics on the cloud overlap over the TP (27 ℃N-39 ℃N;78 ℃E-103 ℃E) region (Fig. 1a).

2.1 Satellite datasets

Radar signals can penetrate the optically thick cloud layers that attenuate lidar signals, but lidar signals may sense the optically thin hydrometeor layers that are below the detection threshold of radar signals. Thus, with the unique complementary capabilities of the CPR on CloudSat and the CALIOP on the CALIPSO, the 2B-GEOPROF-LIDAR dataset produces the most accurate descriptions of the locations of the hydrometeor layers in the atmosphere on the global scale (Mace and Zhang, 2014). In this dataset, every CloudSat profile includes 125 height layers (e.g., vertical bin), and the "*CloudFraction*" parameter reports the fraction of the lidar volume within each radar vertical bin that contains hydrometeors (Mace et al., 2009; Mace and Zhang, 2014). Several previous studies have identified a cloudy atmospheric bin based on different thresholds of the lidar-identified cloud fraction, including a 99% (Barker, 2008; Di Giuseppe and Tompkins, 2015) or a 50% threshold (Haladay and Stephens, 2009; Verlinden et al., 2011). Here, a threshold of 99% is used in our study. Due to the significant attenuation of lidar signals to the optically thick layers, this parameter fails to provide the "*CloudFraction*" for optically thick layers. Thus, we also use the radar information (i.e., cloud "*LayerBase*" and "*LayerTop*" fields) from the aforementioned dataset to construct the complete two-dimensional cloud mask (See Fig. 1b). It is noting that the 2B-GEOPROF-LIDAR dataset does not distinguish cloud and precipitation, therefore any bias in our results caused by precipitation can't be removed in current analysis. Besides the 2B-GEOPROF-LIDAR dataset, the ECMWF-AUX dataset (Partain, 2004), which is an intermediate dataset that are the ancillary ECMWF state variables interpolated across each CloudSat CPR bin, are also used to provide the pressure and height information of each vertical bin in the cloud mask profile. The vertical and horizontal resolutions of these products are 240 m and 1.1 km, respectively. To avoid sunlight scattering contamination to lidar

observation and minimize surface contamination of the CPR, we only use the nighttime datasets above 1 km over the TP surface in our analysis.

2.2 Meteorological reanalysis dataset

The 6-hourly ERA-Interim reanalysis with a grid resolution of 0.25 °× 0.25 °(Dee et al., 2011), is used to characterize the atmospheric thermodynamic and dynamic states over the TP. For each cloud mask profile in the 2B-GEOPROF-LIDAR, the vertical profiles of the zonal wind *u*, meridional wind *v*, relative humidity *rh*, specific humidity *sh* and atmospheric temperature *T* closest to the cloud profiles in both space and time are extracted and further interpolated vertically to match the vertical bins of the cloud mask profile. Following Di Giuseppe and Tompkins (2015), the *u* and *v* winds at every vertical bin are then projected onto the satellite overpass track, being averaged in the along-track direction for all profiles in the selected CloudSat data segment to derive the scene-average, along-track horizontal wind *V*. Here, we define the wind shear $dV/dz_{i,j}$ between the layers *i* and *j*, as follows:

$$dV/dz_{i,j} = \frac{\max\{V_i;V_j\} - \min\{V_i;V_j\}}{D_{i,j}}, \quad (2)$$

where V_i and V_j are the horizontal winds at layers *i* and *j*, respectively, and $D_{i,j}$ is the layer separation distance. The derived wind shear will be used to calculate the cloud overlap parameter. For the CloudSat overpass track (Fig. 1a), Di Giuseppe and Tompkins (2015) indicated that the cross-track shear of the zonal wind *u* has little statistical significance.

Similarly to the wind shear, we calculate the vertical gradient of the saturated equivalent potential temperature ($\partial\theta_{es}/\partial z_{i,j}$) between the same two layers to quantify the dependence of the cloud overlap on the degree of the conditional instability of the moist convection. Here,

$$\theta_{es} = \theta \exp(\frac{L_v r_s}{C_p T})$$

$$\theta = T(\frac{1000}{p})^{0.286}, L_v = 2.5\times10^6 - 2323\times(T-273.16) \quad (3)$$

$$r_{s=} \frac{sh}{rh\times(1-sh)}$$

where θ is the potential temperature, L_v is the latent heat of vaporization, r_s is the saturation mixing ratio, C_p is the specific heat capacity at a constant pressure, and T is the atmospheric temperature. The smaller the $\partial\theta_{es}/\partial z_{i,j}$, the more unstable the atmosphere. Furthermore, the scene-averaged vertical velocity at 500 hPa is extracted from the ERA-Interim reanalysis to analyze the impact of vertical motion on cloud overlap. The positive values are for the updraft, and negative values are for the subsidence.

2.3 The overlap parameter and dependence on the spatial scale

Previous studies have shown that the overlap parameter α and decorrelation length L are sensitive to the spatial scale of the GCM's grid box (Hogan and Illingworth, 2000; Oreopoulos and Khairoutdinov, 2003; Oreopoulos and Norris, 2011; Pincus et al., 2005). For example, Hogan and Illingworth (2000) found that cloud overlap parameter tends to increase with decreasing spatial and temporal resolution (i.e., increasing vertical and horizontal grid scales) of GCMs.

To examine the dependence of overlap parameter over TP on the spatial scale, each CloudSat orbit over the TP region is divided into segments with different horizontal lengths including 25, 50, 100 and 200 km. For convenience, this length is referred to as the spatial scale of the GCM's grid box. Fig.1b shows an example of cloud mask over the TP region from the 2B-GEOPROF-lidar dataset. This cloud mask includes eight, four, two and one segments, corresponding to the horizontal resolution of 25, 50, 100 and 200 km, respectively. Given the threshold of 99% for cloud fraction, the segment-average cloud cover profile of each segment is first derived. Here, it is important to emphasize that cloud fraction and cloud cover are different variables in our study. The "*Cloud fraction*" reports the fraction of lidar volumes in each radar vertical bin that contains hydrometeors and is used to identify a cloudy atmospheric

bin based on the chosen threshold, which is 99% in this paper. When averaging in the along-track direction for all cloud fraction profiles in a selected CloudSat data segment, we derive the segment-average cloud covers profile, which represents the percentage of clouds in a given spatial scale and certain height. Then, the vertical overlap between any two atmospheric layers in this profile is calculated when the cloud covers (C_i and C_j) of both layers exceed 0. Layers are analyzed in pairs and no 'double-counting'. If two cloud layers have the same separation distance but different altitudes, they will be categorized into the same statistic group. Following Hogan and Illingworth (2000) and Di Giuseppe and Tompkins (2015), we consider the nonadjacent layers to be a continuous cloud pair when all layers between them are classified as cloud layers. Otherwise, these layers are classified as a non-continuous cloud pair (Hogan and Illingworth, 2000; Di Giuseppe and Tompkins, 2015).

Based on the definitions of different overlap assumptions and overlap parameter α in the introduction section, Figs.1c and 1d show an example of the observed and calculated segment-average cloud covers profile based on maximum and random assumptions, and corresponding cloud overlap parameter of continuous cloud pair for 25, 50, 100 and 200 km spatial scale in given cloud mask sample (Fig. 1b). It is clear that the observed and calculated cloud covers and corresponding overlap parameters tend to increase as the spatial scale increases. Meantime, the observed cloud covers tend to transform from the maximum to random overlap assumption with increasing layer separations.

By collecting 4 years of cloud sample from the 2B-GEOPROF-LIDAR dataset, Figs.2a and 2b further show the dependence of α on the layer separation and its sensitivity to the spatial scale for both non-continuous and continuous cloud pairs. Many studies have used ground- and space-based radar to examine the validity of the random overlap assumption for the vertically non-continuous clouds (Hogan and Illingworth, 2000; Mace at al., 2002; Naud et al., 2008; Di Giuseppe and Tompkins, 2015). Fig.2a shows that the degree of cloud overlap of the non-continuous clouds over the TP region is lower than the random overlap, especially when the layer separation is smaller than 2km. Given the spatial scale of 50 km, almost all the α

-values are negative and fall between -0.25 and -0.05. Thus, the total cloud cover would still slightly be underestimated for non-continuous cloud pairs by using the random overlap assumption. Assuming a cloud layer separation of less than 9 km, α for non-continuous cloud pairs increases as the spatial scale increases (e.g., from 25 km to 200 km). For a continuous cloud pair (Fig. 2b), α decrease from 0.95 to 0 with an increasing separation. Meantime, a slight dependence of α on the spatial scale is also observed for continuous cloud pairs when they are separated by a distance of about 1 km to 4 km. This indicates that a maximum overlap is slightly more common for a larger horizontal domain, which is consistent with previous studies (Hogan and Illingworth, 2000; Oreopoulos and Khairoutdinov, 2003; Oreopoulos and Norris, 2011).

2.4 Selection of thresholds for cloud cover and spatial scale

About the dependence of α on the spatial scale, Tompkins and Di Giuseppe (2015) theorized that some overcast or single cloud layers would be removed from the samples when the spatial scale is smaller than the cloud system scale, thus biasing α and its decorrelation length. Given a spatial scale of 50 km, the ratio of the spatial scale to the cloud system scale decreases strongly from the equator to the poles because many of the frontal cloud systems of the middle and high latitudes are larger than the convective cloud systems over the tropics. Ultimately, the corresponding bias of α would increase with latitude. Thus, regional atmospheric models should account for the typical cloud system scales in their parameterization schemes when using a fixed horizontal resolution.

Fig. 2c depicts the probability distribution functions (PDFs) of the horizontal scales of the along-track cloud systems at different heights over the TP region. Here, the horizontal scale of a cloud system at a given height along the CALIPSO/CloudSat track is determined by calculating the number of continuous cloud profiles (N) at a given height. Using a 1.1 km along-track resolution for the CPR measurements, the along-track scale (S) of a cloud system is $S=N\times1.1$ km (Zhang et al., 2014; Li et al., 2015). It is clear that the probability of a small-scale cloud system decreases with an increasing height (Fig.2c). The mean horizontal scale of 59.2 km for a cloud system at

a height of 15 km is almost twelve times greater than that (i.e., 4.6 km) at a height of 2 km. For TP region, we can see that the horizontal scales of cloud system below 10 km are smaller than the spatial scale of 50 km, thus we apply the spatial scale of 50 km to perform the following analysis although this scale would still result in significant errors in α at higher atmospheric heights (e.g., 15km) where cloud has large horizontal scale.

In addition, to further reduce the sensitivity of α to the spatial scale caused by data truncation, we also apply a simple data filter following Tompkins and Di Giuseppe (2015) so that only atmospheric layers with segment-average cloud covers smaller than a given threshold of 50% are retained. As stated by Tompkins and Di Giuseppe (2015), data might still be truncated with this filter, but the sensitivity of the results to the spatial scale should largely be reduced. By limiting the spatial scale (50 km) and upper limit of cloud cover (50%), the number of available cloud layer-pair samples is still at least one million, thus ensuring statistical significance. Fig. 2d shows the variations of cloud sample numbers and the cumulative percentages with cloud layer separations for both non-continuous and continuous clouds at a given spatial scale of 50km. It shows that the cumulative proportion of cloud samples significantly increases with layer separation. For the continuous cloud, the cumulative percentage accounts for 90% of all samples when layer separation is smaller than 4 km. Given the 1.1 km along-track resolution of the CPR measurements and a spatial scale of 50 km (that is, about 50 CloudSat profiles), the each cloudy CloudSat profile has a cloud cover about 2% (Di Giuseppe and Tompkins, 2015).

3. Monthly and zonal variations of overlap parameter for continuous clouds

Figure 3a shows the monthly variations in α for the continuous cloud-pair based on pentad-average over the TP. In Fig.3a, the maximum continuous cloud layer separation gradually increases from January (approximately 6 km) to August (beyond 8 km) and then gradually decreases, indicating that cloud systems over TP during summer are thicker than other seasons due to frequent strong convective motions. When the cloud layer separation is less than 1 km, the overlap parameter α has little monthly variation and is always large (even beyond 0.7). However, the monthly

variation of α becomes manifest with a layer separation larger than 1 km. For a 2-km cloud separation, e.g., α reaches its maximum of 0.45 in August and a minimum of 0.1 in February (see Fig. 3d). For a separation of 3 km, α is generally lower but has the similar monthly variation to those seen for a 2-km separation. By checking the negative value of α in Fig.3a, it is clear that even random overlap assumption could underestimate the total cloud cover between two cloud layers with large separation during all seasons except summer. These cloud overlap features may be associated with the unique topographical forcing and corresponding thermo-dynamic and dynamic environment of the TP. In summer, the TP is usually considered as an atmospheric heat source or "air pump" due to its higher surface temperature compared with surrounding regions at the same altitude (Wu et al., 2015). Additionally, a humid and warm air intrudes from the South Asia monsoon area into the lower atmosphere over the TP would intensify the atmospheric instability of moist convection when combined with the enhanced surface heating (Taniguchi and Koike, 2008). This process further promotes the transportation of water vapor into high altitudes and favors the development of convective clouds. Indeed, satellite observations have indicated that cumulus prevails over the TP during the summer (Wang et al., 2014; Li and Zhang, 2016).

Noting a small horizontal scale of cumulus, a 50 km-spatial scale from CloudSat should not bias α estimate too much in our study. However, previous studies have pointed out that precipitation may bias the cloud overlap statistics toward maximum overlap (Mace et al., 2009; Di Giuseppe and Tompkins, 2015). Present studies did not eliminate the influence of precipitation on the overlap parameter. The overlap parameter α would become smaller if the samples with precipitation are removed from the analysis. The feature may be even more obvious during summer due to more frequent precipitation over TP during this season (Yan et al., 2016). The seasonal variation of α was also found at different ground sites (Mace and Benson-Troth, 2002; Naud et al., 2008). For example, Oreopoulos and Norris (2011) indicated that clouds tend to be more random in the winter and most maximum during the summer. In fact, these overlap properties are associated with cloud system scale, which is

dominated by dynamical situation (Tompkins and Di Giuseppe, 2015).

Figures 3b and 3c show the monthly variations in pentad-averaged conditional instability of the moisture convection ($\partial\theta_{es}/\partial z$) and the wind shear ($dV/dz$) for the continuous cloud-pairs over the TP, respectively. The $\partial\theta_{es}/\partial z$ and dV/dz both exhibit obvious monthly variations for all cloud-layer separations. The atmospheric stability and wind shear gradually decrease from January to August and then steadily increase (see Figs. 3c, 3d, 3e and 3f). From Fig.3c, we see that adjacent atmospheric layers during May to September tend to be more unstable and have weak wind shear. These atmospheric states favor the development of clouds and result in maximum overlap between cloud layers. During other month (e.g., December), clouds also tend to follow the maximum overlap more although adjacent atmospheric layers are stable with large $\partial\theta_{es}/\partial z$ and dV/dz. It might be the case that vertical velocities might be large because of extratropical cyclones or other baroclinic instability. With the layer separation increases, atmospheric layers become more stable and then favor random overlap, especially during summer season. These results verify that a more unstable atmosphere tends to favor a maximum overlap over a random one, as shown in previous studies (Mace and Benson-Troth, 2002; Naud et al., 2008). Note that Figs. 3d and 3f might reveal an inconsistency between the wind shear and atmospheric stability. For example, we can see that the wind shear for a 2-km layer distance is greater than that for a 3-km distance, but the atmosphere is also more unstable. This inconsistency is probably because two cloud layers with the same separation but at different altitudes are sorted into the same statistical group. Or, it is also quite possible that other large scale forcings might influence the overlap. In addition, we find the monthly variations in pentad-averaged vertical velocity (ω) at 500 hPa (see Figs.3g and 3h) are also consistent with the monthly cycle of α. It means that vigorous ascent tends to favor maximum overlap. This result agrees well with the previous studies (Naud et al., 2008).

Figure 4 shows the zonal variations of α, $\partial\theta_{es}/\partial z$, dV/dz and ω over the TP.

Figs. 4a and 4b indicate that α is larger in the south part of the TP and smaller to the north. This is mainly because the atmospheric instability over the southern part of the TP enhances the convective activity (Fujinami and Yasunari, 2001). Due to the weakening of the monsoon and the blocking by topography, less water vapor may reach the northern part, and thus fewer clouds from there (You et al., 2014). Compared with the southern TP, the stability and wind shear are both larger over the northern part, especially for those cloud layers with large separation (e.g, >2km). This meteorological condition will result in more frequent negative α, indicating that random overlap assumption used in models would underestimate the total cloud cover and thus bias the surface radiation over these regions (see Fig.4a). The most significant warming occurring over the Northern part of TP has been attributed to pronounced stratospheric ozone depletion (e.g., Guo and Wang, 2012). However, a more recent study indicates that the accelerated warming trend over the Tibetan Plateau may be due to the rapid cloud cover increases at nighttime over the northern Tibetan Plateau and the sunshine duration increase in the daytime over the southern Tibetan Plateau (Duan and Xiao, 2015). Therefore an accurate representation of total cloud cover and its relations to atmospheric thermodynamic and dynamic conditions in models is critically important to the understanding of the TP rapid warming. Although it is still difficult for models to capture the cloud overlap properties, especially for those cloud layers with large separation over north TP, our results confirm that α is well related with wind shear and instability. However, the zonal variation of α is inconsistent with the variation of vertical velocity (see Figs. 4g and 4h).

4. Sensitivity of α on the meteorological conditions and its parameterization

To facilitate the parameterization of α for cases of continuous clouds, we further investigate the sensitivity of α on the different meteorological conditions. Here, each meteorological factor over the TP region is grouped into one of four bins as follows. The four bins for $\partial\theta_{es}/\partial z$ are $\partial\theta_{es}/\partial z > 5$ K/km, $2.5 < \partial\theta_{es}/\partial z < 5$ K/km, $0 < \partial\theta_{es}/\partial z < 2.5$ K/km and $\partial\theta_{es}/\partial z < 0$ K/km. For wind shear, the four bins

are $dV/dz < 0.5\,\mathrm{m \cdot s^{-1}/km}$, $0.5 < dV/dz < 2\,\mathrm{m \cdot s^{-1}/km}$, $2 < dV/dz < 3.5\,\mathrm{m \cdot s^{-1}/km}$ and $dV/dz > 3.5\,\mathrm{m \cdot s^{-1}/km}$. For vertical velocity, the four bins are $\omega < -40$ hPa/day, $-40 < \omega < 0$ hPa/day, $0 < \omega < 40$ hPa/day and $\omega > 40$ hPa/day. These groupings ensure that a statistically significant number of samples fall within each bin (i.e., at least one hundred thousand samples per bin). In addition, Li et al. (2015) indicated that the overlap properties between different cloud types are obvious different but the most significant components of the global climate system. Although current study doesn't include the information of cloud type, the sensitivity of α on the meteorological parameters in our analysis actually exhibit the effects of cloud types due to different cloud types with same layer separation possibly take place in distinct wind shear and stability conditions.

Figure 5 illustrates the sensitivity of α to wind shear, instability and vertical velocity at given upper limit of cloud cover (50%) and spatial scale (50 km) for the continuous clouds. Since the cloud samples with layer separations smaller than 3.5 km account for 90% of all samples for continuous clouds, we only present the results for layer distances smaller than 3.5 km. Naud et al. (2008) tested the sensitivity of α to wind shear at three sites and found that wind shear slightly affects α when the layer distance is larger than 2 km. In a recent study, Di Giuseppe and Tompkins (2015) demonstrated the important effect of wind shear on the global cloud overlap by using a combination of the CloudSat-CALIPSO cloud data and the ECMWF reanalysis dataset. Our results along with previous studies suggest that the cloud overlap strongly depends on atmospheric conditions, but their relationship displays some variability, in particular spatially and seasonally. The effect of the atmospheric stability on cloud overlap may be more important over convective regions (e.g., the intertropical convergence zone and TP during summer season) while the effect of wind shear may be dominant over the mid-latitudes. Besides the wind shear and instability, some studies also tested the sensitivity of the overlap parameter to the large-scale vertical velocity. For example, Naud et al. (2008) indicated that vertical velocities in the tropics are not captured in the reanalysis dataset when convection occurs, thus they

only discussed the impact of vertical velocity on the cloud overlap parameter over the mid-latitude and found that vigorous ascent tends to favor maximum overlap. Fig.5c shows that vertical velocity at 500hPa has an effect to cloud overlap parameter. However, by combining the effects of wind shear, instability and vertical velocity in parameterizations of decorrelation length scale *L*, we find that this scheme doesn't show better superiority than the scheme which only includes the wind shear and instability. Thus, we only parameterize decorrelation length scale *L* as a function of the wind shear and atmospheric stability in current study.

Here, we derive the decorrelation length scale *L* values (km) from the least squares exponential fit to the original α curve at given wind shear and instability bin. We parameterize *L* as a function of wind shear or both wind shear and atmospheric instability based on a (multiple) linear regression. The regression formula of *L* can be written as:

$$L = L_{\alpha} - b1\frac{\partial \theta_{es}}{dz} - b2\frac{dV}{dz}$$
$$or \qquad (4)$$
$$L = L_{\alpha 1} - c1\frac{dV}{dz}$$

Here, L_{α} , $L_{\alpha 1}$, *b1* ,*b2*,and *c1* are the fitting parameters. Table 1 lists several parameterization schemes for the decorrelation length scale *L*. The scheme with wind shear from Di Giuseppe and Tompkins (2015) using the global CloudSat-CALIPSO cloud data and ECMWF reanalysis dataset is shown for a comparison. Di Giuseppe and Tompkins (2015) discussed the uncertainties from fitting methods and calculation of wind shear. Related to the observational orbit, the impact of cross-track wind shear is neglected in our study, which would exclude many large wind shear associated with jet structures (Di Giuseppe and Tompkins, 2015). The parameterization scheme of Shonk et al. (2010) is also shown in Table 1, which is an empirical linear relationship between *L* and latitude based on CloudSat and CALIPSO data. Our parameterization schemes in terms of wind shear or both wind shear and instability are given in Table 1. Note that the R-squared values (R^2) for our wind shear and wind shear-instability

schemes are 0.88 and 0.96, respectively.

After deriving the regression formula of decorrelation length scale *L*, we re-apply it to all continuous cloud samples and retrieve the *L* and corresponding α based on the formula: $\alpha = e^{-D/L}$ and cloud layer separation. Finally, retrieved overlap parameter α is used to calculate the cloud cover between any two cloud layers by using the Equ.(1) and definitions of random and maximum overlap assumptions. Figure 6 presents the monthly difference between calculated and observed cloud covers using various overlap parameterization schemes. It is seen that the maximum and random overlap assumptions result in large cloud cover biases, especially for layer separations greater than 1 km for maximum overlap and less than 2 km for random overlap where the bias exceeds 5%. Compared with random and maximum assumptions, the differences of cloud over caused by other schemes are small and range from -3% to 3%. In addition the wind shear scheme and the wind shear-instability scheme from the present studies overall show less biases than other schemes. However, several points still need further attention. First, the wind shear scheme from Di Giuseppe and Tompkins (2015) significantly underestimates the cloud cover for layer separations above 1 km (e.g., reach 3%). This large bias may be because it is based on the global CloudSat-CALIPSO measurements and ECMWF reanalysis dataset for a short period (January-July 2008); as such, some obvious regional or seasonal cloud overlap properties are easily obscured by global averaging. Furthermore the role of atmospheric stability was not considered in this scheme. However, the scheme from Di Giuseppe and Tompkins (2015) causes little bias for layer separations below 1 km. This is because this scheme retrieves much larger *L* and overlap parameter values than other schemes. An interesting finding is that Shonk/latitude scheme leads to comparable bias with new schemes from this study. The bias is even smaller for Shonk/latitude scheme when the layer separation is below 1 km. In fact, Fig.5 has demonstrated that the sensitivity of α to wind shear and instability is rather weak when cloud layers are very close. Compared with our wind shear scheme, our wind shear-instability scheme further combines the impact of atmospheric instability and

has a relatively lower bias at large layer separations with higher R-squared values (R^2=0.96).

Fig.7 shows the zonal difference between calculated and observed cloud covers for the aforementioned schemes. The differences of cloud cover caused by different overlap schemes are distinguishable. Similar with Fig.6, the maximum and random overlap assumptions still result in the most prominent cloud cover biases (exceed ± 5%) at most of the layer separations. Compared with our wind shear scheme and wind shear-instability schemes, the scheme from Di Giuseppe and Tompkins (2015) and latitude scheme from Shonk et al. (2010) cause relatively obvious underestimation of total cloud cover when cloud layer separations exceed 1 km, especially for scheme from Di Giuseppe and Tompkins (2015) (bias reach -3%). Only if cloud layer separations are smaller than 1 km, these two schemes produce better cloud cover simulation than our schemes. In summary, these results indicate that new parameterization (that is, our wind shear-instability scheme) of decorrelation length scale, which includes the effects of both wind shear and atmospheric stability on cloud overlap, may improve the prediction of cloud cover over TP.

5. Conclusions and discussion

The Tibetan Plateau has experienced a rapid warming over the past three decades. Previous studies suggests that the change in cloud cover may explain different temperature trends in the daytime and nighttime over the TP (Duan and Wu, 2006; Kang et al., 2010). Indeed, many studies have verified that annual and seasonal total cloud amounts have declined over TP (e.g., Yang et al., 2012; You et al., 2014), leading to the increase of absorbed solar radiation and the increase of surface air temperatures (Duan and Xiao, 2015).

To accurately simulate the total cloud cover and its impact on the radiative energy budget, climate models need to reliably represent the cloud vertical overlap, which has received less attention than necessary because of the limited regional cloud observations. In this study, we analyze 4 years (2007–2010) of data over the Tibetan Plateau from the CloudSat 2B-GEOPROF-LIDAR dataset and the ECMWF-AUX dataset and along with the ERA-Interim daily 6-hourly reanalysis. It is confirmed that

the continuous cloud layers tend to have maximum overlap at small separation but gradually become randomly overlapped with an increase of the layer separation. Focusing on the continuous cloud layers, we evaluate the effects of the meteorological conditions on the cloud overlap. It is found that the unstable atmospheric stratification with a weak wind shear over the TP would tend to favor maximum overlap (that is, greater α values), agreeing well with previous studies. We parameterize the decorrelation length scale *L*, a parameter that is used to characterize the transition from the maximum to random overlap assumption, as a function of the wind shear and atmospheric stability. Compared with other parameterizations, this new scheme improves the prediction of cloud cover over TP when cloud layers separations are greater than 1km. Although the parameterization method derived in our study focuses only on the TP, our results suggest that the parameterization of the decorrelation length scale *L* by considering multiple thermodynamic and dynamic factors and microphysical effects (e.g., precipitation) has the potential to improve the model-simulated total cloud covers.

In a recent study, Di Giuseppe and Tompkins (2015) applied the wind shear-dependent decorrelation length scale in the ECMWF Integrated Forecasting System. They found that the impact of wind shear-dependent parameterization on radiative budget calculation is comparable in magnitude to that of latitude-dependent scheme of Shonk et al. (2010). Our results also show that latitude-dependent scheme has similar bias of cloud cover relative to the new scheme developed in this study. Although these results can't suggest which of the scheme is superior, the scheme based on factor the meteorological factors has some potential advantages. For example, cloud overlap parameter is significantly controlled by atmospheric thermodynamic and dynamic conditions, therefore the long-term variations of meteorological factors are bound to affect the trend of cloud overlap and the resulting in total cloud cover and radiation budget. Indeed, a recent study has shown that rapid warming and an increase of atmospheric instability over the TP leads to more frequent deep clouds, which are responsible for the reduction of solar radiation over the TP (Yang et al., 2012). By using surface observations over 71 stations, some studies

verified that annual and seasonal total cloud covers have declined during 1961-2005 (Duan and Wu, 2006; You et al., 2014). However, whether such variations of total cloud cover are linked with the changes of degree of cloud overlap over the TP are still unclear. Thus, more efforts are needed to reasonably evaluate the impact of cloud overlap on the total cloud cover variations over these sensitive areas of climatic change (e.g., Tibetan Plateau and Arctic).

Competing interests. The authors declare that they have no conflict of interest.

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Table 1. Parameterizations of decorrelation scale length L from the exponential fit as a function of atmospheric stability $\partial\theta_{es}/\partial z$, wind shear dV/dz or latitude Φ

Scheme	description	decorrelation length scale L
Wind shear (Di Giuseppe and Tompkins, 2015)	Random/Maximum, only wind shear	$L = 4.4 - 0.45 \times \frac{dV}{dz}$
Wind shear (this study)	Random/Maximum, only wind shear	$L = 2.19 - 0.14 \times \frac{dV}{dz}$
Wind shear-instability (this study)	Random/Maximum, wind shear and instability	$L = 2.18 - 0.09 \times \frac{dV}{dz} - 0.15 \times \frac{\partial\theta_{es}}{dz}$
Latitude (Shonk et al., 2010)	Random/Maximum, only latitude	$L = 2.899 - 0.02759 \times \lvert \Phi \rvert$

Figure Captions

Figure 1. (a) CloudSat overpass tracks (blue line: daytime; red line: nighttime) over the Tibetan Plateau (27 ℃N-39 ℃N; 78 ℃E-103 ℃E); (b) A sample of CloudSat 2B-GEOPROF-LIDAR cloud mask product along the ground track of 200km (white color: cloud fraction>99%; light blue: 0<cloud fraction<99%; deep blue: clear sky; orange color: surface). (c) The observed and calculated segment-average cloud covers profiles based on maximum and random assumptions for different spatial scales and given cloud mask sample in Fig. 1b. (d) The corresponding cloud overlap parameters of continuous cloud pair for 25, 50, 100 and 200 km spatial scale. Note that the observations below 1 km over the TP surface have been removed.

Figure 2. The dependence of α on the layer separation and its sensitivity to the spatial scale for (a) non-continuous and (b) continuous cloud pairs; the error bars

correspond to ± 3 standard error; (c) The probability distribution functions (PDFs) of the along-track horizontal scales of cloud system at different height over TP region; (d) The variations of cloud sample numbers and the cumulative percentages with cloud layer separations for both non-continuous and continuous clouds at a given spatial scale of 50km. The cumulative percentages represent the proportions of cloud sample below corresponding layer separation to all samples.

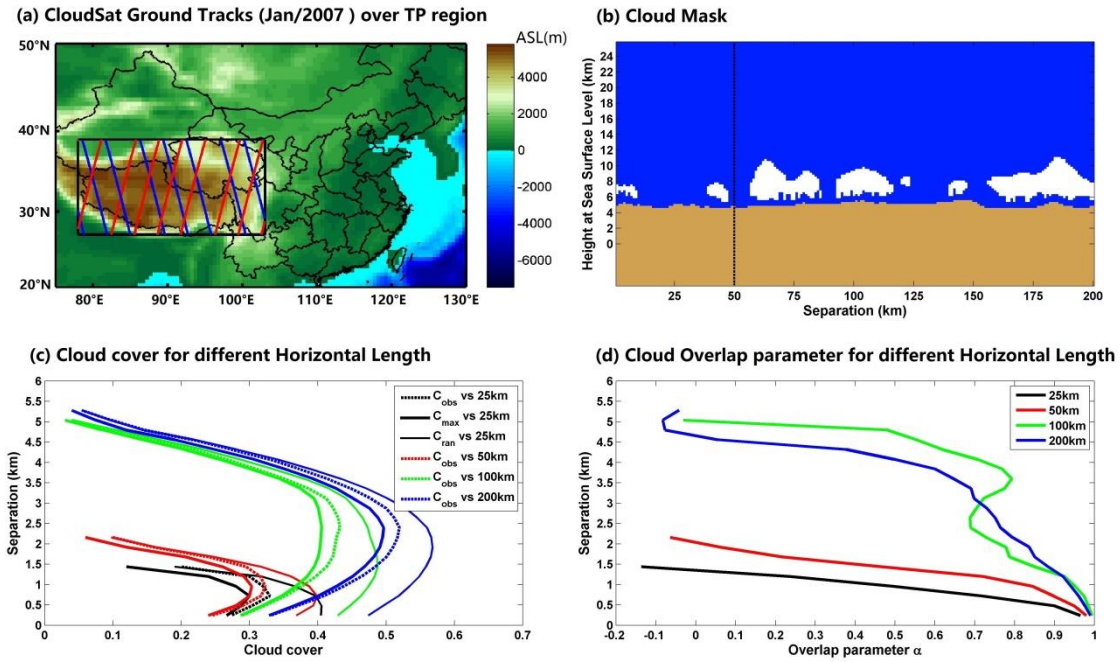
Figure 3. The monthly variations of the pentad-averaged (a) cloud overlap parameter, α, (c) conditional instability to moist convection, $\partial\theta_{es}/\partial z$, (e) wind shear, dV/dz, (g) and vertical velocity at 500 hPa, ω for the continuous clouds over the TP ; The monthly variations of the pentad-averaged (b) α, (d) $\partial\theta_{es}/\partial z$, (f) dV/dz and (h) ω for the continuous clouds for the layer separation of 2 km (red) and 3km (black).

Figure 4. The zonal variations of the (a) α, (c) $\partial\theta_{es}/\partial z$, (e) dV/dz, and (g) ω for the continuous clouds over the TP ; The zonal variations of the (b) α, (d) $\partial\theta_{es}/\partial z$, (f) dV/dz and (h) ω for the continuous clouds for the layer separation of 2 km (red) and 3km (black).

Figure 5. The sensitivities of median overlap parameter α to (a) wind shear, (b) instability and (c) vertical velocity at 500 hPa at given upper limit of cloud cover (50%) and spatial scale (50 km) for the continuous clouds. The error bars correspond to ± 3 standard error.

Figure 6. The monthly differences in cloud cover between calculation and observation for different schemes (see the Table 1) and its dependence on the layer separation.

Figure 7. The zonal differences in cloud cover between calculation and observation for different schemes (see the Table 1) and its dependence on the layer separation.



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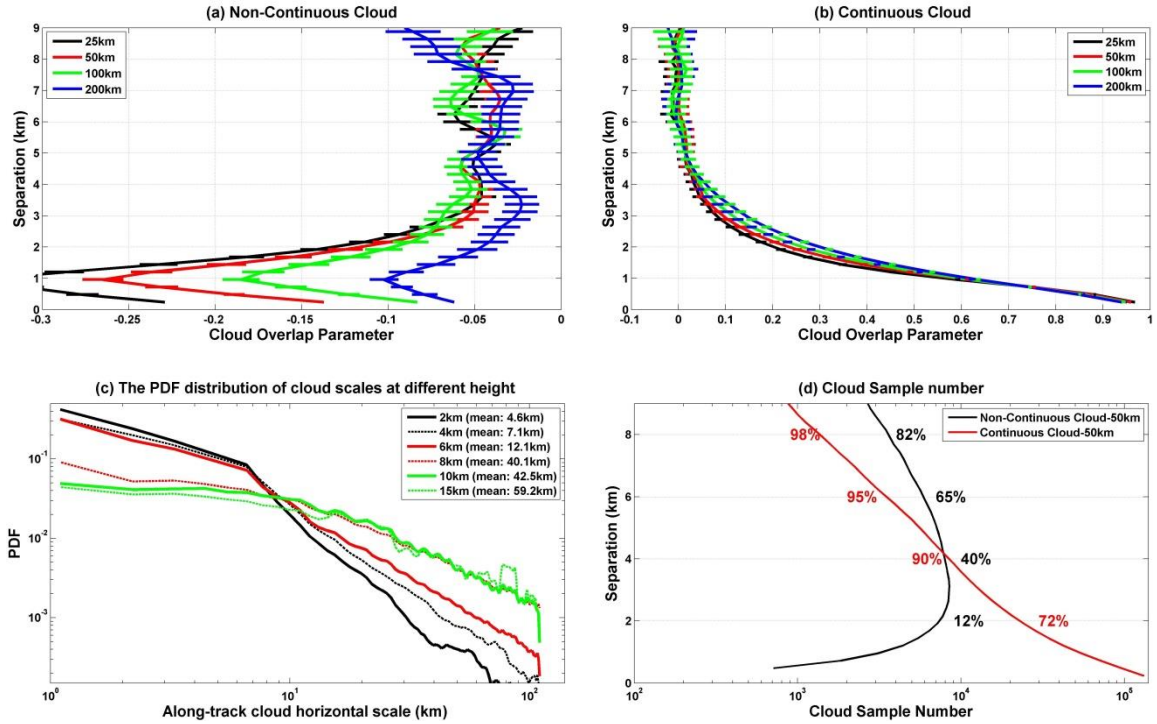

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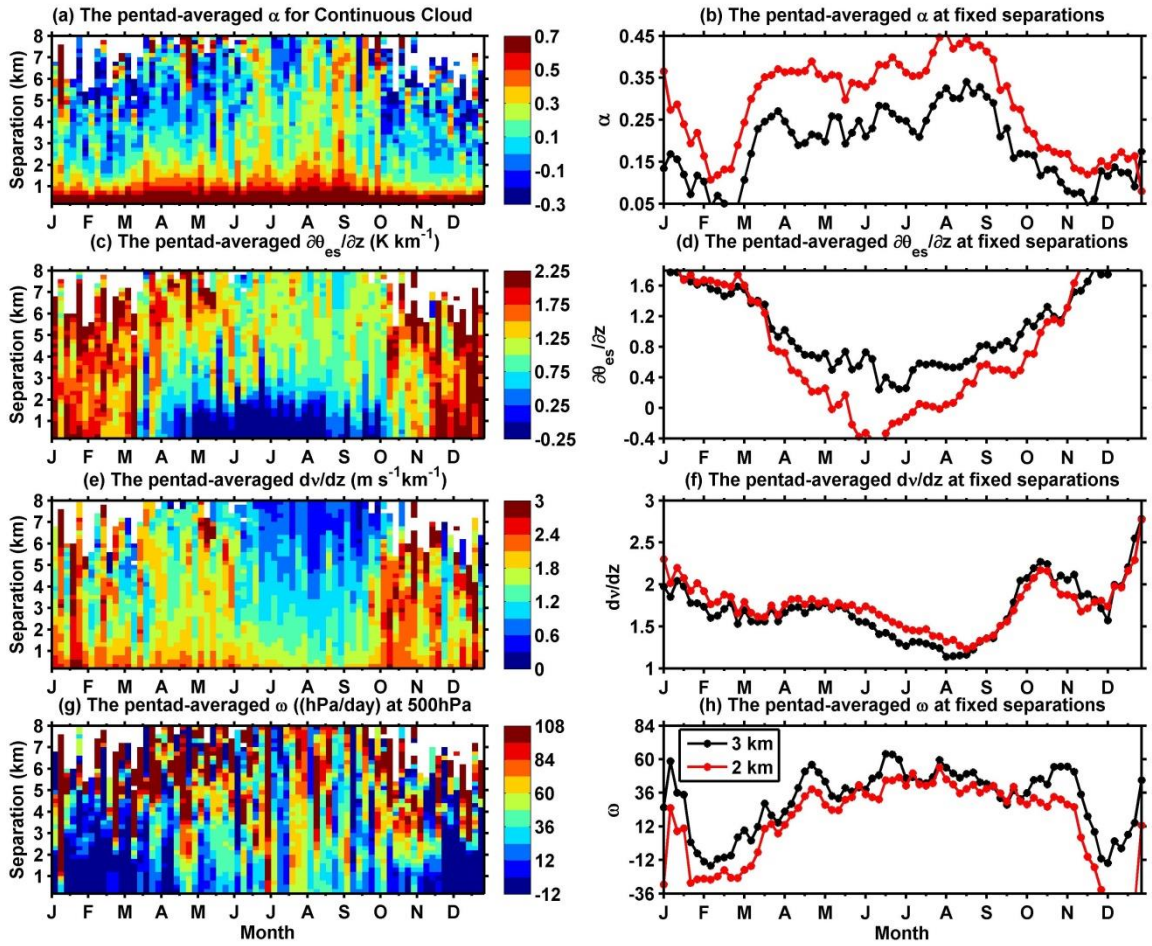

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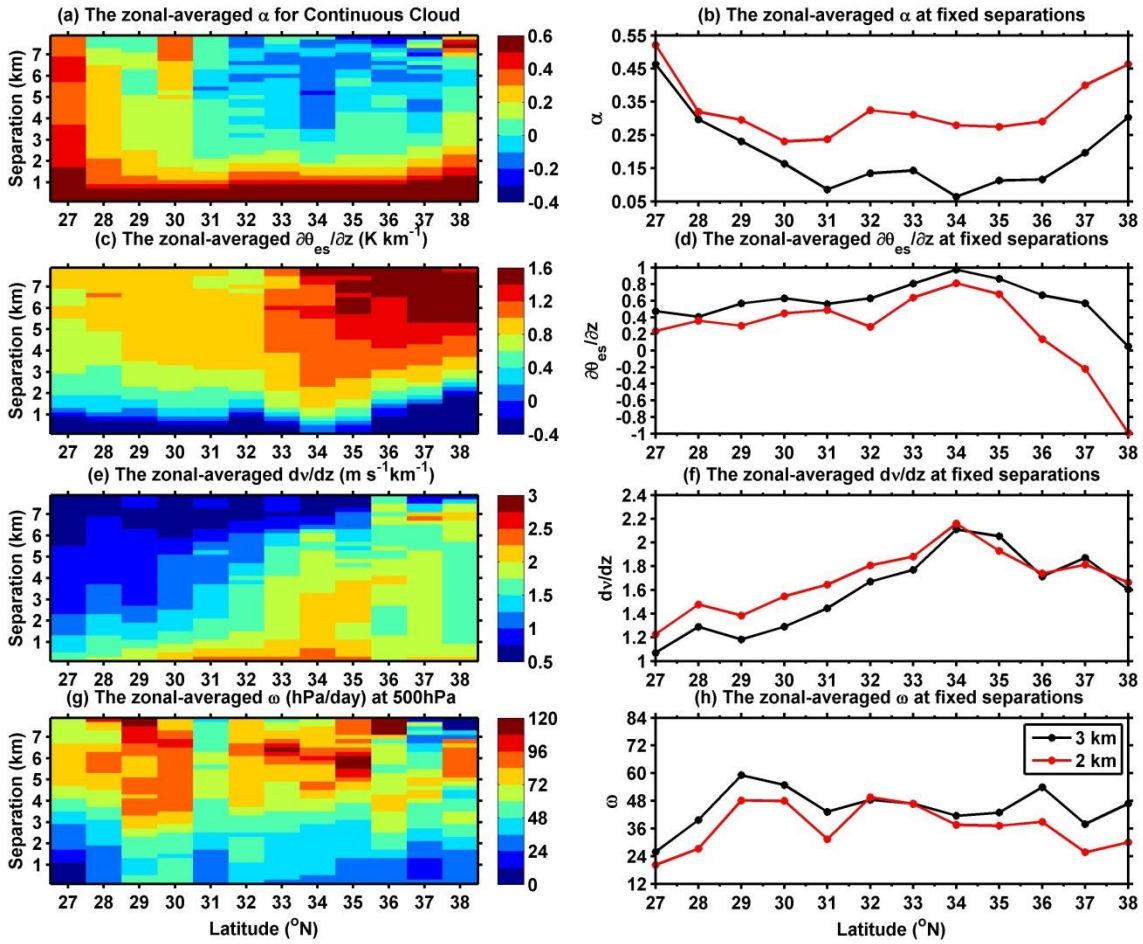

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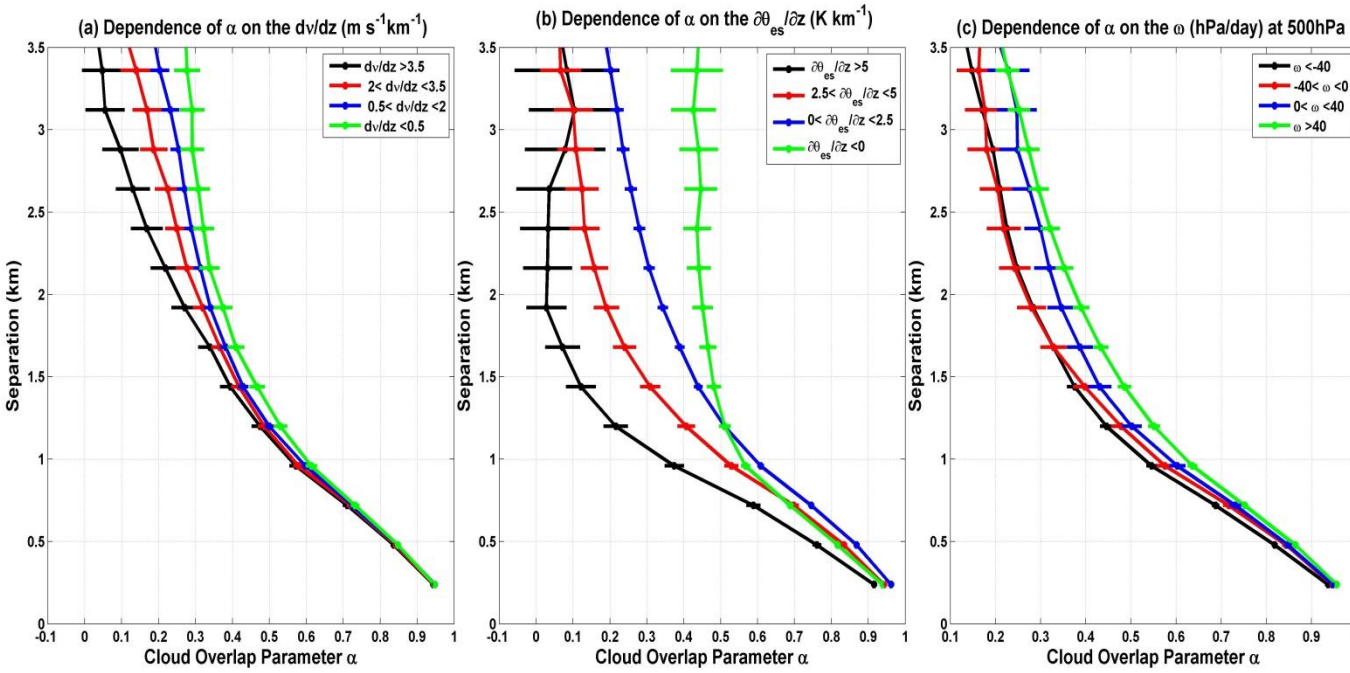

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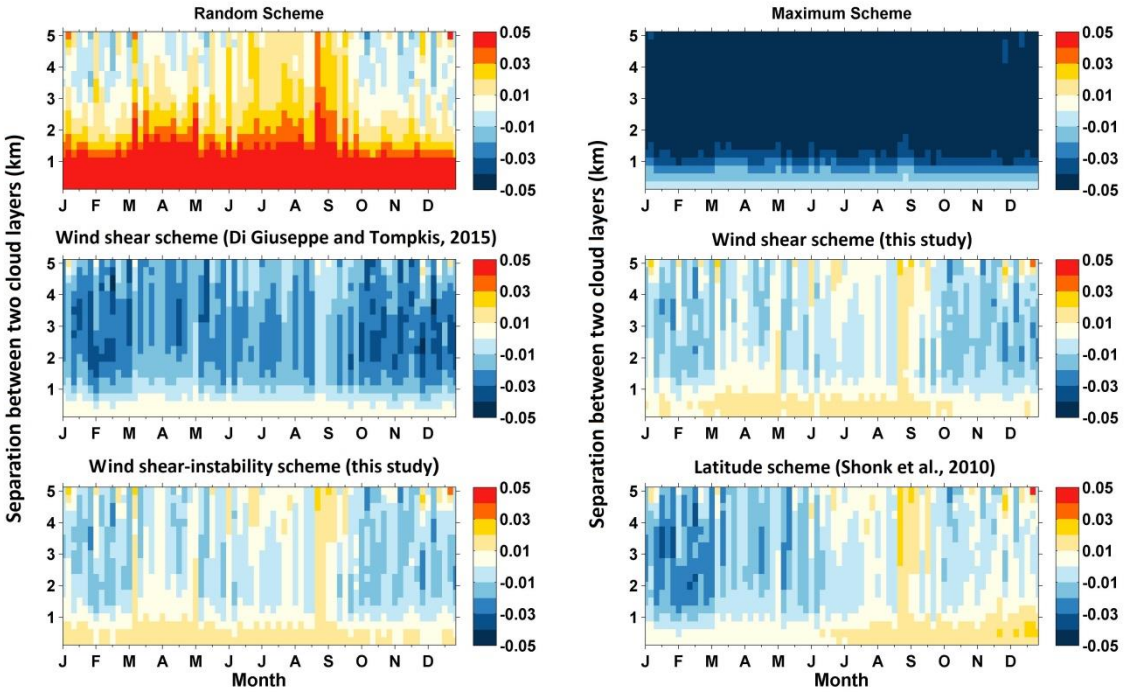

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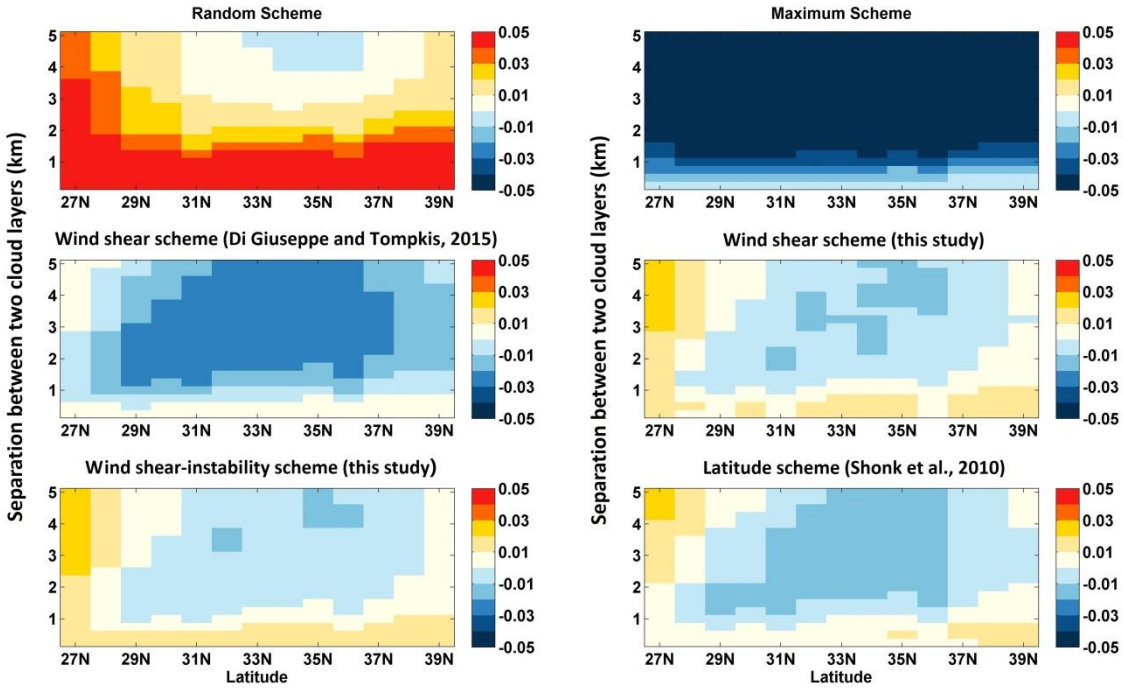

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