Manuscript No: acp-2017-55
Journal: ACP
The revised manuscript entitled "**Direct radiative effects of dust aerosols emitted from the Tibetan Plateau on the East Asian summer monsoon – a regional climate model simulation**" by Hui Sun, Xiaodong Liu, and Zaitao Pan.

We are very thankful to the two reviewers for their insightful and constructive comments. We believe that their comments helped us highlight some critical aspects of the paper and improve the quality of our work. We have addressed all the concerns in a point-by-point manner. In the following, the underlined *italic* texts are reviewer's comments and normal (font) texts are our responses. The bold texts have been inserted to new version of our manuscript.

Referee #1 (Comments to Author):

This is an interesting study which investigates the direct radiative effects of dust aerosols emitted from the Tibetan Plateau on the East Asian summer monsoon with a regional aerosol-climate model. In general, it is well written and structured and there are original model results presented and discussed. However there are a number of major comments that have to be taken into consideration before acceptance of the manuscript for publication.

We now add more analyses and discussions which are summarized below:

(1) compared simulated dust AOD with pure dust observation from CALIPSO in Figure 6.

(2) added anomaly of geopotential height in Figure 11 and updated the figure of precipitation.

(3) carried out two new sensitive experiments to isolate the effects of changed land cover alone on the results, and discussed uncertainty brought by the method used in the manuscript.

(4) compared the aerosol-induced signal on the meteorological fields with that from the model's internal variability.

(5) added more discussions throughout the manuscript to clarify uncertainty of our simulations, and updated the references.

Comments:

1. Please discuss briefly what is the added value of using a regional climate model instead of global climate model to study the impact of aerosols on climate.

In the revised version, we added the following statements to characterize advantages of using

RCM instead of GCM to study the impact of aerosols on climate (Page 11, Line 3–10).

It is very beneficial to study the impact of aerosols on climate using a RCM instead of a coarse-resolution global climate model (GCM). The GCMs tend to systematically underestimate dust aerosol concentration, presumably due to their lower spatial resolution (Tegen et al., 2002; Zhang et al., 2009). The RCM simulated dust concentration, on the other hand, was closer to the observed magnitude compared with the results of global models (Sun et al., 2012). For example, previous studies showed that high-resolution RCMs had better capabilities than GCMs in simulating the effect of aerosols on Asian monsoon (Zhou and Yu, 2006; Gao et al., 2008; Ji et al., 2011). A high-resolution regional model is especially needed to capture subtle characteristics in the areas of complex terrain (Ji et al., 2011).

2. In the discussion section the authors should also comment on the limitations of using a regional climate model to study the impact of aerosols on climate. For example, could the authors comment if an RCM which is actually forced by lateral boundary conditions of a GCM or reanalysis can be able to provide the adequate spatial coverage for the development of atmospheric circulation feedbacks over a limited area.

Yes, RCM's high horizontal resolution comes at the expanse of being covering a limited area. Although RCM's domain is relatively small in a global sense, the atmospheric circulation away for the domain still enter the domain via lateral boundary condition (BC). Given a properly designed BC and reasonably sized domain, RCM simulations should be able to provide adequate spatial coverage for atmospheric circulation to develop and exert feedback in the domain. The reasonably simulated in wind and temperature pattern compared with the observations did suggest the external forcing passed through the lateral boundary well and developed properly within the RCM domain. At your suggestions, the following elaboration and caution are added to Discussion Section (Page 11, Line 10–16).

However, limited area RCM naturally cannot fully account for external forcing remote from the domain of interest although the lateral boundary conditions allow large-scale features to propagate into the domain. Our domain size (9600 km × 640 km) is reasonably large enough so that the weather and climate systems can have adequate spatial extent to develop within the domain, as attested by reasonably validation of wind pattern, temperature field, and precipitation (Section 3.1). Cautions should be exercised, however,

that results from regional simulations could be somewhat domain-size dependent quantitatively although main results should not be affected.

3. An issue that it is not discussed at all is if the aerosol induced signal on the meteorological fields is higher than the model's internal variability. Did the authors carried out some sensitivity experiments to investigate this important issue? I think at least a few comments on this issue are necessary. This is also a part of the limitations in these simulations.

We deeply appreciate this great comment and agree that the model's internal variability could be large, so we compared the standard deviation of summer surface temperature and precipitation in CON with signals induced by the dust effects (CON minus SEN) during heavy dust years (Figure A1). It seems that the signal induced by the dust is much greater than the standard deviations. Therefore, the dust effects in our simulation is significant. We mentioned this in the revised version as below (Page 10, Line 17–22).

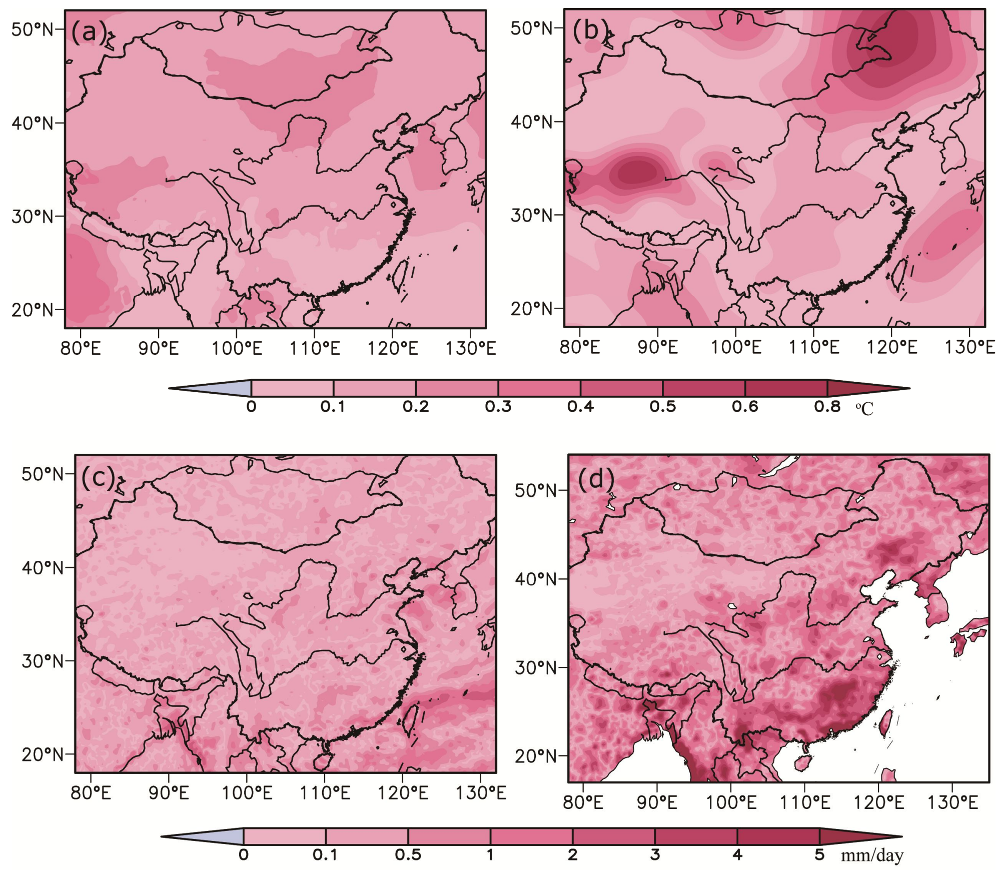

Figure A1: Standard deviation of summer surface temperature (a) and precipitation (c) in CON, and differences (absolute value) of summer surface temperature (b) and precipitation (d) between CON and SEN during the heavy dust years.

It is worth mentioning that the model's internal variability could influence the results; so we compared the standard deviation of summer surface temperature and precipitation in CON with the signal induced by the dust effects (CON minus SEN) during the heavy dust years. The signal induced by the dust is much greater than the standard deviations (figures not shown). Therefore, the dust effects reported in our simulation is significant in the heavy dust years, but the cooling over central India in the light dust years may be caused by the model's internal variability.

4. Page 3, lines 4-7: There are a number of other recent published studies that have looked the effect of aerosols on climate using RegCM e.g. Das et al., Clim. Dyn., 2015 and Das et al., TAC, 2016 for Asia, Zanis et al., Clim.Res. 2012 for Europe, Ji et al., Clim. Dyn., 2015 and Komkoua et al., Int. J. Clim., 2017 for Africa.

Thanks. We have cited these articles in the revised version (Page 3, Line 8–9).

5. There is a recent study by Tsikerdekis et al., ACP (2017) testing a newly implemented 12-bin approach for RegCM which is also compared with the default RegCM4 4-bin approach used in the RegCM simulations of this work.

Thanks. We read and cited the reference in the revised version as below (Page 11, Line 21–23).

A recent study by Tsikerdekis et al. (2017) demonstrated that simulated dust load and induced radiation change are sensitive to the dust particle size division in the model; so further sensitivity experiments using more dust size bins would be worthwhile.

6. Please clarify if the RegCM simulations in this work use only dust aerosols or other aerosols as well (such as anthropogenic or marine aerosols). To my understanding the simulations include only dust particles. Hence the comparison of modelled dust AOD with AOD from satellite (MISR) and ground based (AERONET) measurements is not one to one comparison since these observations include all types of aerosols. Mind though that there are available pure dust satellite products from CALIPSO (see e.g. Amiridis et al., ACP, 2013 and Marinou et al., ACP, 2017).

Only dust aerosols included in our simulation, and we have clarified this point in the revised version as below (Page 4, Line 26–27).

In order to isolate the effect of dust aerosols, only dust aerosols are included in our simulations, without considering other aerosols (such as anthropogenic or marine aerosols).

We agree that these comparisons are imperfect, and are grateful for reminding us availability of the pure dust observation from CALIPSO. We added a section to compare the simulated dust AOD with those of observed by CALIPSO, and the comparison is good (Figure 6). Accordingly, we updated the data description in Section 2 (Page 5, Line 18–22), and added following figure and texts in the revised version (Page 7, Line 1–11).

3.1.4 Simulated and CALIPSO-observed dust AOD comparison

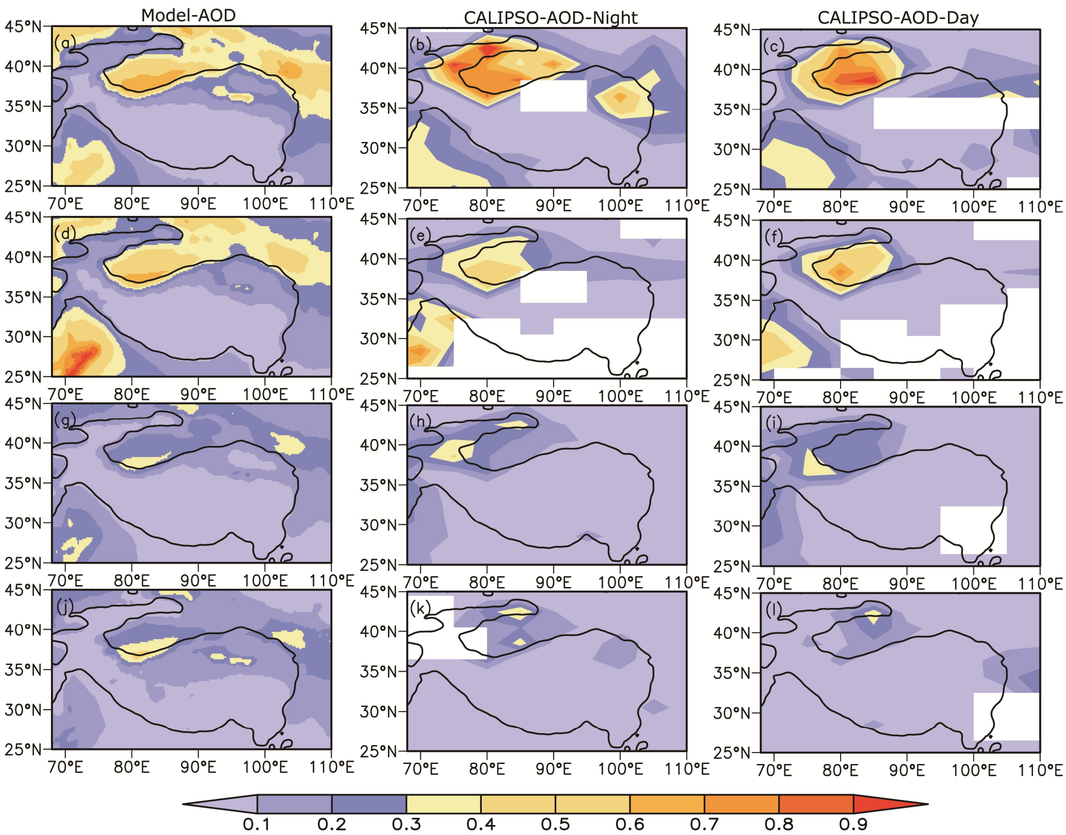

Figure 6: Spatial distribution of the dust AOD simulated by the control experiment (left panels) and observed by CALIPSO at night (middle) and during daytime (right panels) averaged in (a, b, c) spring, (d, e, f) summer, (g, h, i) autumn and (j, k, l) winter during the time period 2007–2009.

While MISR and AERONET data contain all types aerosols including those anthropogenic ones, the CALIPSO observation sole devotes to dust aerosols. Figure 6 shows that the simulated seasonal variation, center positions and magnitude of dust AOD are very

consistent with those observed by CALIPSO during day and night. Both simulations and observations not only showed that dust AOD increased in spring and summer and decreased in autumn and winter, but also captured three maximum centers of dust AOD in Taklimakan, the Great Indian Desert and Qaidam Basin located in the northern TP in spring. The simulated center values were still high in summer. Besides, it is very interesting to noted that the observed dust AOD in the Qaidam Basin is higher at night than that during daytime (Figure 6b and 6c), which implied that dust activities in the TP may be more prominent at night. This unusual feature could cause dust radiative effects in the TP is very different from those of in other locations, which is further discussed in subsequent section of 3.4.1.

7. The authors mention that in order to eliminate the dust emission in the Tibetan Plateau they replaced the land cover types of these areas with the nearby vegetated types. This change could stop the dust emission but will also change the surface albedo which itself could have an impact on radiation budget, temperature and circulation. In other words if it is like this the results do not show simply the effect of eliminating the emission of dust particles but also the effect of land cover type and albedo change. Please comment on this important issue.

We deeply appreciate this great comment. We agree that our method is imperfect and uncertainty exists in the results. To address your concerns, we carried out two additional sensitive experiments to evaluate effects of the changed land cover. Dust cycle in the two experiments was turned off, but the land cover was changed to become similar to the modification in SEN. The differences between the two experiments only included effects of the changed land cover. The results showed that change (from desert to vegetated land) brought about 0.4 °C warming (Figure A2 and Figure A3), and this warming effect is weaker than the combined cooling effects (–0.6°C) induced by the dust aerosols and the changed land cover. Besides, we have noted that Li and Xue (2010) demonstrated that land cover change from vegetated land to bare ground (mainly desert) in the TP decreased the radiation absorbed by the surface and resulted in weaker surface thermal effects, which means that effect of land cover change from desert to vegetated land may also cause warming effects. This is opposite to the cooling effects induced by the dust aerosols over the TP in our simulation, and it may partly offset the dust aerosols induced cooling effects. However, our results showed that the cooling signal was not changed, and it may even be underestimated in the heavy and light dust years. Therefore, actual cooling should be stronger than the simulated. We will try a new better method in our future work. We mentioned the uncertainty brought by the method and discussed the influences on the results in the revised version as below (Page 10, Line 22–34, and Page 11, Line 1–2).

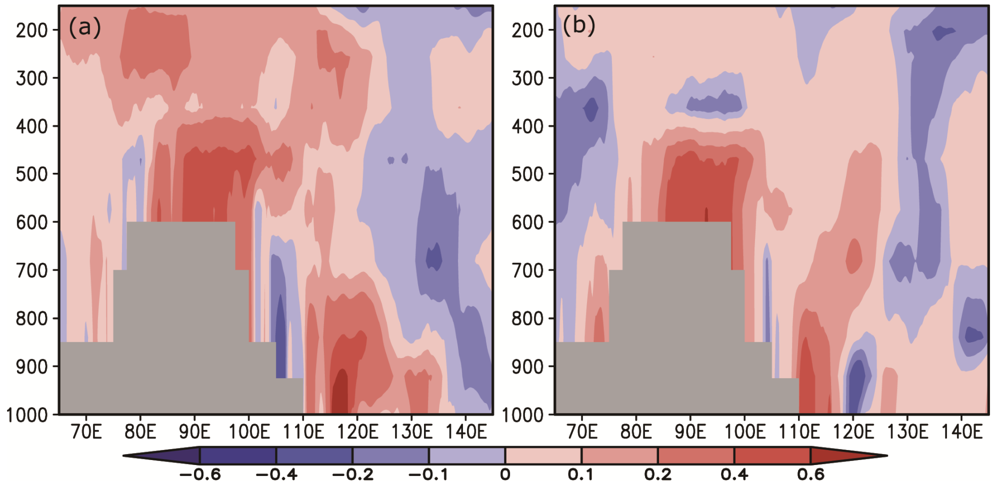

Figure A2: Longitudional-height cross-section of air temperature anomaly induced by the land cover changed averaged over 32–36°N in summer in heavy (a) and light (b) dust years.

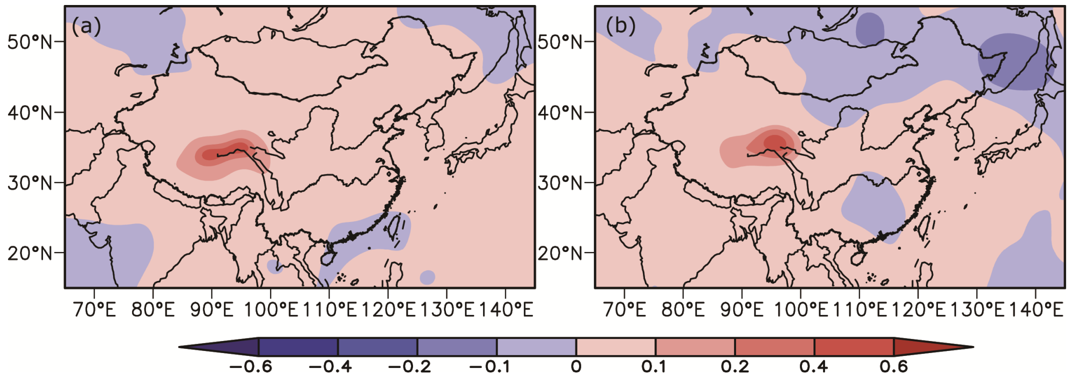

Figure A3: Summer surface temperature change induced by the land cover change in (a) heavy and (b) light dust years.

Besides, the results could also include the role of changed land cover in addition to the role of dust aerosols because turning off dust emission in the TP was through modifying underlying surface types. Hence, we carried out two additional sensitive experiments to isolate effects of the changed land cover alone. The dust cycle in the two experiments was turned off, but the land cover was changed to one similar to the modification in SEN. The differences between them only included effects of the change in land cover. The results showed that the change (from desert to vegetated land) brought about 0.4 °C warming (figures not shown), and this warming effects is weaker than the combined cooling (–0.6°C) induced by the dust aerosols and the changed land cover together. Besides, it is interesting to note that Li and Xue (2010) had demonstrated that the land cover change from vegetated land to bare ground (mainly desert) in the TP decreased the radiation absorbed

by the surface and resulted in weaker surface thermal effects, which means that the effect of land cover change from desert to vegetated land may also cause warming effects. This is opposite to the cooling effect induced by the dust aerosols over the TP in our simulation, and the warming may partly offset the dust aerosols-induced cooling effect. However, our results showed that the signal of cooling effects was not changed, although it may even be underestimated. Therefore, actual cooling should be stronger than the simulated value. Hence, the reported dust effects also need be evaluated by using a refined way in the future.

8. The authors mention in Section 3.2 that "the dust aerosol increases and decreases over the TP as the EASM index weakens and enhances, respectively". Please could provide some discussion on the physical explanation for this anti-correlation. Also provide some short description for the EASM index used.

We added some short description for the EASM index in Section 3.2 in the revised version as below (Page 7, Line 14–15).

This index measures the intensity of the southerly wind to the east of TP in lower troposphere above East Asia.

We added following schematic diagram and texts for physical explanation of the anti-correlation in the revised version (Page 9, Line 16–25).

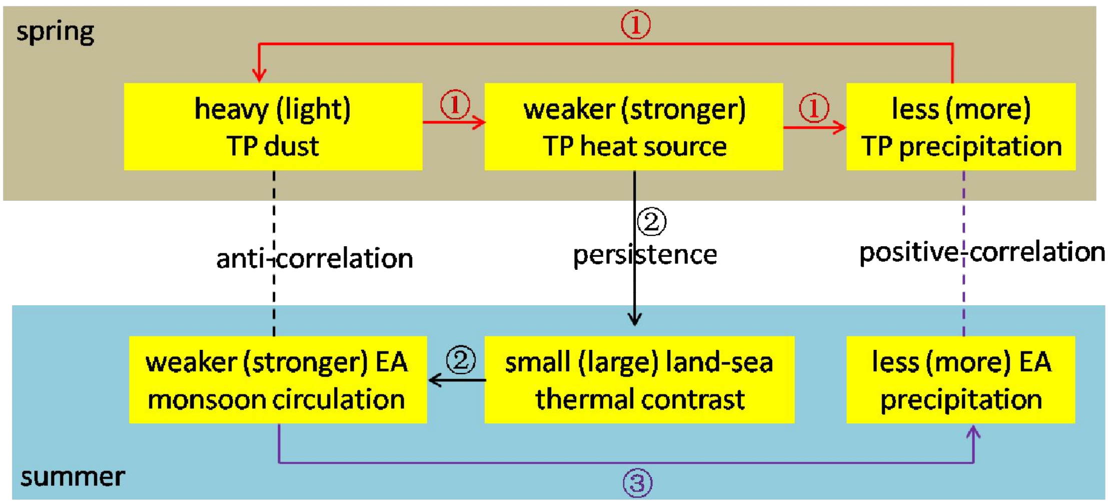

Figure 13 Schematic diagram showing the relation of dust aerosols emitted from the TP in spring with EASM precipitation

The spring dust aerosols from the TP have a close relation with EASM. Although the cause-effect relationship is not immediately clear, the following processes are proposed as a possible mechanism of this relation based on the results in our simulation (Fig. 13). Firstly, increasing (decreasing) in dust aerosols over the TP in the heavy (light) dust years in spring can weaken (enhance) the TP heat source and thus reduce (increase) precipitation over the TP. Reduction (increase) in precipitation over the TP can also further enhance (diminish) dust emission over the TP. Secondly, the weakened (enhanced) TP heat source can persist from spring to summer and shrink (expand) the land-sea thermal contrast and thus weaken (enhance) the EASM. Therefore, the change of dust over the TP has an anti-correlation with the variation of EASM circulation intensity. Thirdly, weakened (enhanced) monsoon circulation can reduce (increase) precipitation in East Asia. As a result, the precipitation variation of the TP presents a positive-correlation with that of EASM.

9. The authors point is Section 3.4.1 that the effect of TP aerosols on surface temperature are not limited to the areas that the dust aerosols are locally emitted. So the effect on temperature is not solely due to local radiation imbalance from the presence of dust aerosols but also due to aerosol induced circulation changes. Similar results have been pointed for Asia by Das et al., 2015 as well as in earlier studies by Zanis, 2009 and Zanis et al., (2012) for Europe.

Thanks. We read and cited these references in the revised version as below (Page 8, Line 16–18).

Similar phenomena were reported earlier from Europe for anthropogenic aerosols (Zanis 2009; Zanis et al., 2012) and from South Asia for natural aerosols (Das et al., 2015a).

10. It would be helpful if the authors could also add in Figure 10 the Geopotential Height anomalies with colors to point spatially the anticyclonic circulation anomaly.

Thanks for the suggestion. We updated this figure and rephrased this part. Please see Section 3.4.2 in our revised version (Page 8, Line 20–25).

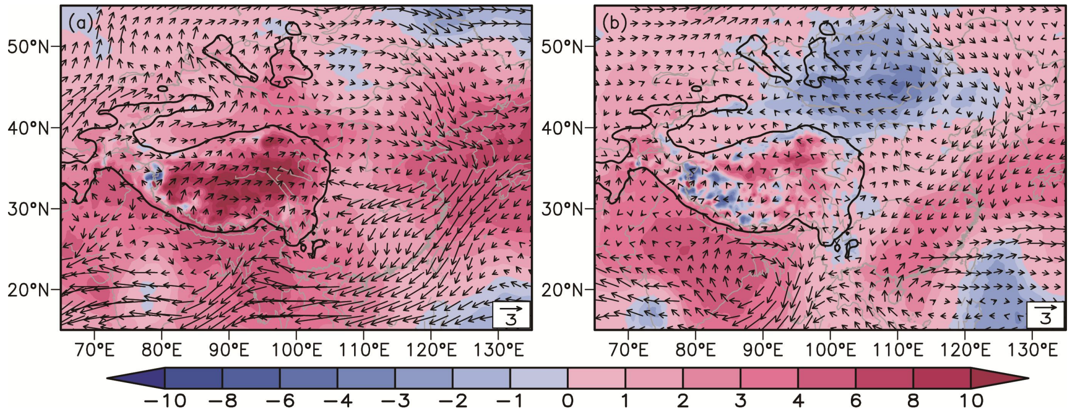

Figure 11: Simulated differences in atmospheric circulation at 850 hPa (vector, m s^{-1}) and geopotential height at 600 hPa (shaded, m) in summer between CON and SEN in (a) heavy and (b) light dust years.

The overall effects of TP aerosols cool the troposphere surrounding the TP (Fig. 10a) and thus the land–sea thermal contrast was reduced by the dust aerosols over the TP. The atmospheric circulation anomaly induced by the dust aerosols emitted over the TP in heavy dust years shows an overall gigantic anticyclonic circulation centered over the TP **with a positive anomaly (>10m) in geopotential height** (Fig. 11a). The northeasterlies that run against the southwesterly monsoon is especially strong over the EASM region, which indicates that the EASM was weakened greatly. The anomaly still existed in the light dust years, but its intensity was much weaker than in the heavy dust years (Fig. 11b)**.**

11. The discussion in Section 3.4.3 for the dust particle effect on precipitation in heavy and light dust years needs more elaboration. Maybe the authors could include a figure for precipitation similar to Figure 9 for temperature.

Thanks for the suggestion. We updated this part with a horizontal distribution figure for precipitation change similar to temperature and rephrased this part. Please see Section 3.4.3 in the revised version (Page 8, Line 26–31, and Page 9, Line 1–3).

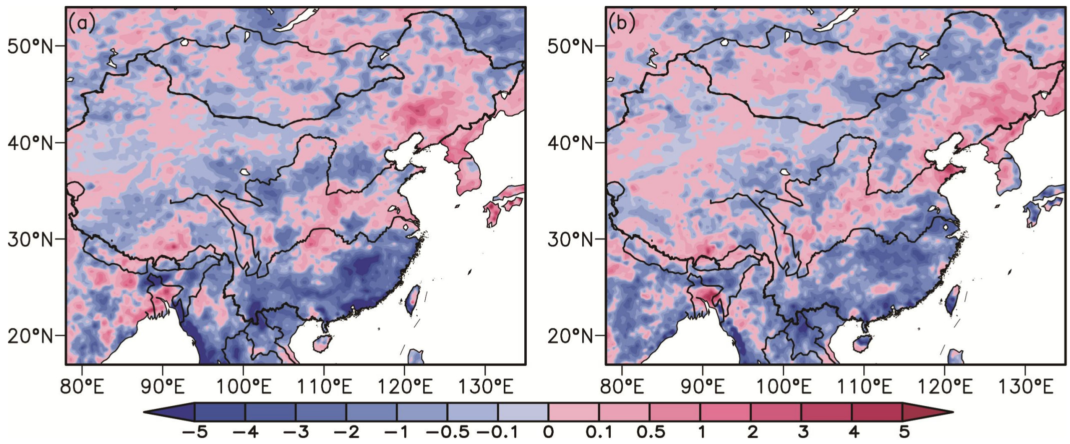

Figure 12: Simulated difference in summer precipitation between CON and SEN in (a) heavy and (b) light dust years.

Figure 12 shows the simulated change in summer precipitation in **East Asia** induced by dust emitted over the TP in heavy and light dust years. The precipitation decreased in **both the southern and the northern monsoon regions** in summer during heavy dust years as a result of weakening EASM (Fig. 11), **and the reduction in the southern monsoon region is greater than that in the northern monsoon region**. The dust aerosols also reduced precipitation in the two monsoon regions in the light dust years. The simulated suppressive effects of the dust aerosols were consistent with previously reported modeling results (Sun et al., 2012; Guo et al., 2015). **Besides, precipitation in the heavy dust years reduced more than that in the light dust years in the TP, which may be suppressed by the enhancement of descending motion induced by the strong cooling effects of dust aerosols over the TP.**

12. The conclusion that Figure 12 shows that the dust aerosols emitted over the TP delay the onset of the EASM is really weak since no uncertainty analysis is implemented and the differences between control and sensitivity experiments are small. I think this statement needs more elaboration and justification.

Yes. Results of this part is not very robust. Since it is off the major focus of this paper, we delete this part in our revised version. We will explore this in the future.

13. Since the work is focusing on summer monsoon season I think that Figure 2 (a, b , c and d) should also refer to the summer season for consistency reasons, similarly to Figures 2e and 2f .

All the figures in Figure 2 are summer average. We rephrased the caption of Figure 2 as below (Page 20, Line 6).

Figure 2: Spatial distribution of (a, b) **summer** surface air temperature (°C) and (c, d) **summer** precipitation (mm day^{-1}) simulated in the control experiment (left) and the CRU observations (right) for 1990–2009. The bottom two panels are wind vectors at 850 hPa simulated in (e) the control experiment and (f) the NCEP–DOE re-analysis during the summer monsoon season (June–August) averaged for 1990–2009.

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Referee #2 (Comments to Author)

General comments: The authors use an RCM in order to investigate the effect of Tibetan Plateau dust sources on the East Asian Summer Monsoon (EASM). They find that removing the desert cells in Tibet reduces precipitation and generally weakens the EASM. The subject of the study is interesting and original, the presentation is clear (but a little lacking in depth) and the flow is smooth. There are a few major points that need to be clarified or otherwise addressed before the paper is accepted.

We now add more analyses and discussions which are summarized below:

(1) compared simulated dust AOD with pure dust observation from CALIPSO in Figure 6.

(2) added anomaly of geopotential height in Figure 11 and updated the figure of precipitation.

(3) carried out two new sensitive experiments to isolate the effects of changed land cover alone on the results, and discussed the uncertainty brought by the method used in the manuscript.

(4) compared the aerosol-induced signal on the meteorological fields with that from the model's internal variability.

(5) added more discussions throughout the manuscript to clarify uncertainty of our simulations, and updated the references.

Specific comments in order of appearance :
p.1, l.24: "dust ... accounts for about half of all the aerosols". By mass? This is not supported from Table 1 of the Chin et al. 2002 reference.

Yes. Dust aerosol has the largest emissions (1500 Tg/yr) and abundance of mass (32.2 mg/m^2) compared to other aerosols (IPCC, 1994). We overstated a little in the first draft. The wording was likely to lead to ambiguity; so we delete these words in our revised version (Page 1, Line 26).

p.5, l.10: The alleged supremacy of MISR over MODIS is justified based on only one paper which itself is based on only one AERONET station. The better agreement of MISR AOD with the specific station cannot be considered representative, see for example Bibi et al., 2015 (http://dx.doi.org/10.1016/j.atmosenv.2015.04.013) who show that MISR compares better with AERONET at two stations, and MODIS at the other two stations of the Indo-Gangetic plains. I wouldn't exclude MODIS from the analysis.

MODIS AOD has a large portion of missing data in Northwest China; so we only used the data from MISR. Our statement was inaccurate and thus we rephrased these sentences as below (Page 5, Line 14–22). We also used dust AOD from CALIPSO to evaluate our simulations (Page 7, Line 1–11).

level-3 monthly mean AOD data during 2000 to 2009 obtained from the Multiangle Imaging Spectroradiometer (MISR) onboard NASA's Earth Observation System Terra satellite (http://www-misr.jpl.nasa.gov/). Since MODIS AOD has a large portion of missing data in Northwest China, MISR was used to evaluate the simulated dust AOD in CON. The effectiveness of the MISR data was investigated by Martonchik et al. (1998, 2004) and Bibi et al. (2015).

p.6, l.20: It would be nice to include some statistics (correlation coefficient, bias, etc) in Figs. 4 and 5

Thanks. We added correlation coefficient in Figure 4 (Page 22) and Figure 5 in the revised version (Page 23).

p.7, l.13 and Fig.8: The widespread aerosol-induced cooling is quite impressive, but also raises questions. In much of the literature the direct radiative effect of dust is predominantly positive (warming) over land areas and becomes negative in specific situations like a large zenith angle (e.g. Quijano et al., 2000, J. Geophys. Res, 105(D10), 12207-12219). Specifically over Tibet, Chen et al. (2006) (reference in manuscript) show net aerosol warming. I would suggest that the authors explore more their derived aerosol cooling and provide information on the reasons behind this behaviour. For example, is the LW cooling from dust particles so much larger than the SW warming? How much less absorbing in SW is Tibetan dust compared to dust from other locations? What are the optical properties of the dust emitted by Tibet?

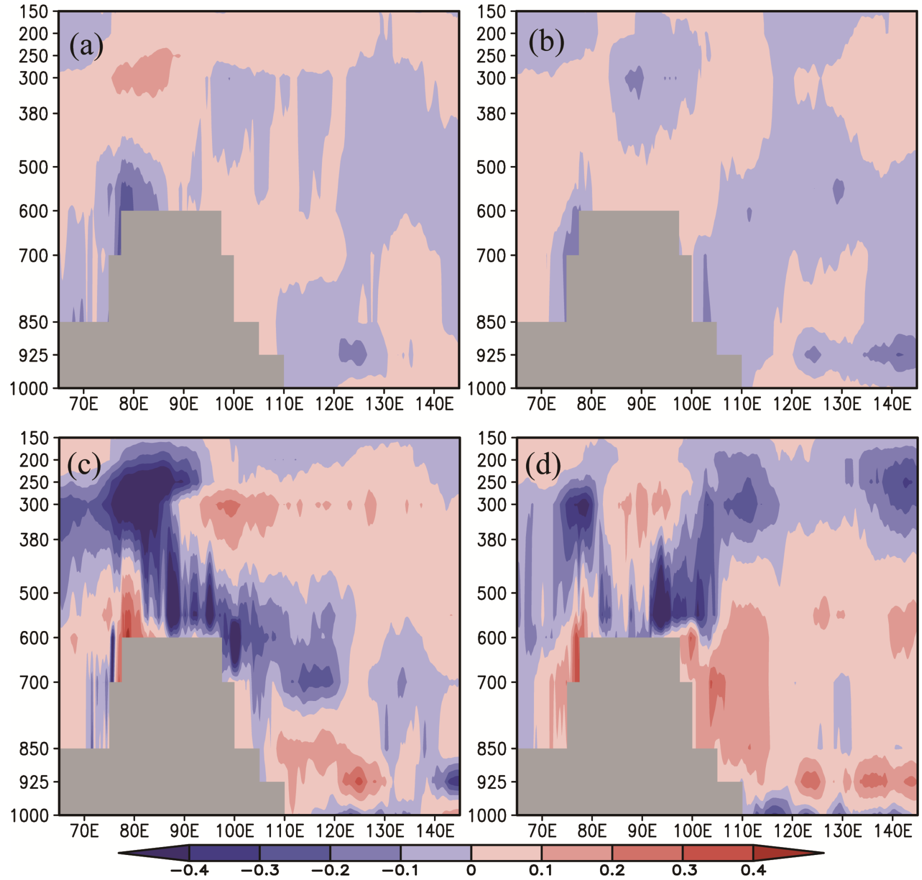

Figure B1 Longitudinal cross-section of the differences between CON and SEN averaged over 32–36°N in summer. (a) and (b): net short-wave heating rate (K day^{-1}). (c) and (d): long-wave cooling rate (K day^{-1}) in heavy (a, c) and light dust years (b, d) respectively.

We deeply appreciate this great suggestion. We agree that direct radiative effect of dust is predominantly positive over land areas in most of previous reports (Zhang et al., 2013; Chen et al., 2013). We further examined differences in the dust AOD between day and night over the TP using the pure dust observation from CALIPSO and found that the observed dust AOD at night is higher than that during daytime over the TP (Figure 6b and 6c), which means that the dust over the TP is more active at night than daytime. Therefore, SW warming is much weaker than LW cooling (Figure B1). We noted the positive (warming) effects reported by Chen et al., (2013), but their results included the effects of other strong absorbing aerosols such as black carbon. Besides, they did not exclude contribution of other dust sources including that from the Taklimakan and the Gurbantunggut Desert to the north of the TP and that from the Great Indian Desert to the south of the TP. The warming effects in their study may be caused by black carbon or the dust aerosols from Taklimakan during daytime, but the cooling effects in our study is mainly caused by the dust aerosols emitted from the TP at night. We added comparison between the simulated and CALIPSO observed dust AOD in the revised version (Page 7, Line 1–11), and discussed the differences (Page 10, Line 7–16) between our simulations and those of simulated by Chen et al., (2013). The following texts and Figures 6 are added in the new version.

3.1.4 Simulated and CALIPSO-observed dust AOD comparison

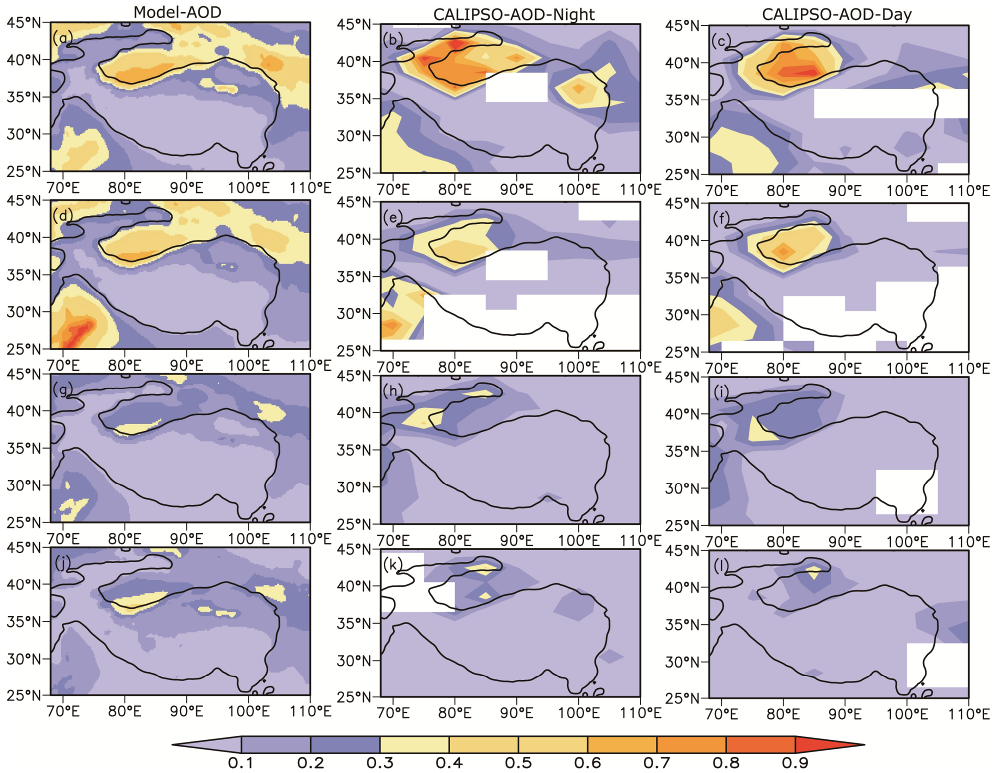

Figure 6: Spatial distribution of the dust AOD simulated by the control experiment (left panels) and observed by CALIPSO at night (middle) and during daytime (right panels) averaged in (a, b, c) spring, (d, e, f) summer, (g, h, i) autumn and (j, k, l) winter during the time period 2007–2009.

While MISR and AERONET data contain all types aerosols including those anthropogenic ones, the CALIPSO observation sole devotes to dust aerosols. Figure 6 shows that the simulated seasonal variation, center positions and magnitude of dust AOD are very consistent with those observed by CALIPSO during day and night. Both simulations and observations not only showed that dust AOD increased in spring and summer and decreased in autumn and winter, but also captured three maximum centers of dust AOD in Taklimakan, the Great Indian Desert and Qaidam Basin located in the northern TP in spring. The simulated center values were still high in summer. Besides, it is very interesting to note that the observed dust AOD in the Qaidam Basin is higher at night than that during daytime (Figure 6b and 6c), which implies that dust activities in the TP may be more prominent at night. This unusual feature could cause dust radiative effects in the TP

is very different from those of in other locations, which is further discussed in the subsequent section of 3.4.1

Dust direct radiative effect is reported predominantly positive (warming) over land areas in most of previous researches (Zhang et al., 2013; Chen et al., 2013), but it can become negative under specific situations like a large zenith angle (Quijano et al., 2000). It is interesting to note that it can also become negative when dust activities are mainly vigorous at night over the TP. The observed dust AOD at night is much higher than those of during daytime over the TP (Figure 6b and 6c); so the SW warming effects is quite weaker than the LW cooling effects (figures not shown). The simulations of Chen et al., (2013) included strong absorbing aerosols such as black carbon and included contribution from other dust sources such as the Taklimakan and the Gurbantunggut Desert to the north of the TP and that from the Great Indian Desert to the south of the TP; thus the warming effects of dust aerosols reported in their study is broader. In contrast, the cooling effects in our study is mainly caused by the dust aerosols emitted from the TP at night.

p.7, l.18: It would be interesting to see why the dust generates this downward motion.

Geopotential heights (at 600 hPa) increased in most of land in East Asia (Figure 11a), and the downward motion induced by the LW cooling of dust over the TP is enhanced (Figure B1 C). The following texts (Page 8, Line 10–11) were added in the discussion of revised version.

The enhanced descending motion was induced by the LW cooling effects of dust over the TP (figures not shown).

Fig.9: How does the Tibetan dust cause cooling over central India only during the light dust years? I'm afraid that using heavy and light dust years introduces aerosol unrelated interannual variability that complicates the picture. It would be much better if it were possible to tweak the dust productivity directly (please see below).

Thanks for pointing this out. Interaction of aerosols with EASM is very complicated, and many factors can influence the monsoon. Contribution of dust aerosol to the monsoon variability may be only a small part. Perhaps the dust effects is significant only in the heavy dust years or monsoon anomaly years. Therefore, we focus on the heavy and light dust years. Because of less dust aerosol in the light dust years, the cooling over central India may be caused by the model's internal variability (Figure B2 a). We explained this in the revised version as below (Page 10, Line 17–22).

It is worth mentioning that the model's internal variability could influence the results; so we compared the standard deviation of summer surface temperature and precipitation in CON with the signal induced by the dust effects (CON minus SEN) during the heavy dust years. The signal induced by the dust is much greater than the standard deviations (figure not shown). Therefore, the dust effect reported in our simulation is significant in the heavy dust years, but the cooling over central India in the light dust years may be caused by the model's internal variability.

p.7, l.27 and Fig 10: As mentioned also by referee #1, the anticyclonic activity might be better visualized through geopotential heights.

Thanks for the suggestion. We updated the figure and rephrased this part, Please see Section 3.4.2 (Page 8, Line 19–25)

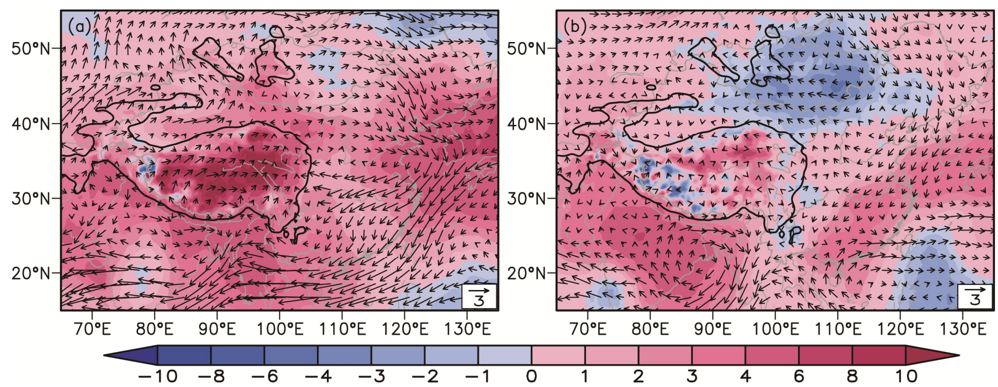

Figure 11: Simulated difference in atmospheric circulation at 850 hPa (vector, m s^{-1}) and geopotential height at 600 hPa (shaded, m) in summer between CON and SEN in (a) heavy and (b) light dust years.

The overall effects of TP aerosols cool the troposphere surrounding the TP (Fig. 10a) and thus the land–sea thermal contrast was reduced by the dust aerosols over the TP. The atmospheric circulation anomaly induced by the dust aerosols emitted over the TP in heavy dust years shows an overall gigantic anticyclonic circulation centered over the TP **with a positive anomaly (>10m) in geopotential height** (Fig. 11a). The northeasterlies that run against the southwesterly monsoon is especially strong over the EASM region, which indicates that the EASM was weakened greatly. The anomaly still existed in the light dust years, but its intensity was much weaker than in the heavy dust years (Fig. 11b).

p.8, l.15 and Fig.12: If I understand correctly, this difference in the EASM onset is rather marginal and probably circumstantial. For example if the value 5.5 were selected instead of 6, then the CON experiment shows earlier onset. The aerosol-induced delay of the EASM does not seem like a robust result.

Yes. Result of this part is not very robust. Since it is off the major focus of this paper we delete this part in our revised version. We will explore this in future.

Section 3.4.3: I think it would be interesting to show the change in precipitation with a Figure similar to Figs. 9 and 10. Also, there is no mention of precipitation changes in the north monsoon region.

Thanks for the suggestion. We updated this part with a new figure for precipitation change similar to temperature and rephrased this part. Precipitation in the northern monsoon regions was also suppressed by the dust. Please see Section 3.4.2 (Page 8, Line 26–31, and Page 9, Line 1–3).

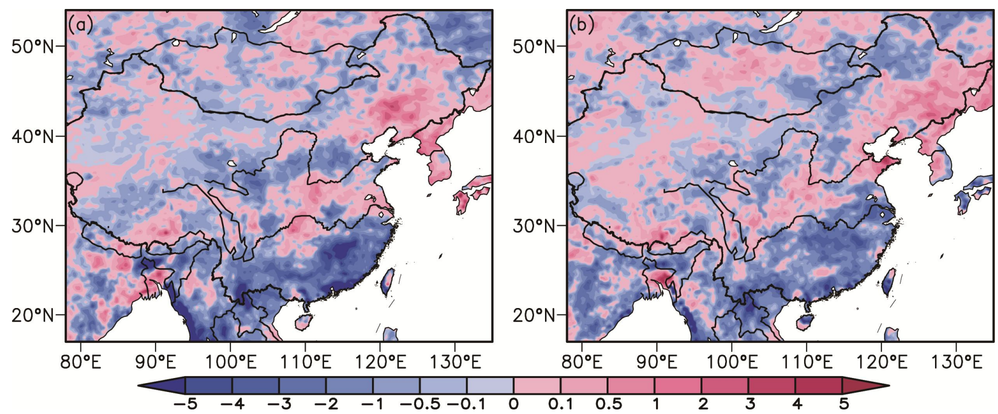

Figure 12: Simulated difference in summer precipitation between CON and SEN in (a) heavy and (b) light dust years.

Figure 12 shows the simulated change in summer precipitation in **East Asia** induced by dust emitted over the TP in heavy and light dust years. The precipitation decreased **in both the southern and the northern monsoon regions** in summer in the heavy dust years as a result of weakening of the EASM (Fig. 11), **and the reduction in the southern monsoon region is greater than that in the northern monsoon region**. The dust aerosols also reduced precipitation in the two monsoon regions in the light dust years. The simulated suppressive effects of the dust aerosol were consistent with previously reported modeling results (Sun et al., 2012; Guo et al., 2015). **Besides, precipitation in the heavy dust years reduced more than that in the light dust years in the TP, which may be suppressed by the enhancement of descending motion induced by the strong cooling effects of dust aerosols over the TP.**

A general remark: The authors focus on heavy and light dust years in order to evaluate the EASM sensitivity to aerosol emissions. Relying only on the heavy/light year classification, the problem retains the interannual variability from irrelevant factors such as the meteorological fields. Instead of (or maybe complementary to) the heavy/light year experiment, I would try reducing or increasing by specific percentages (e.g. 10%- 100% in steps) the dust emission from the surface of Tibet, through modifications in the dust module. Then I would try to present the "climatological" 20-yr average change. I am not experienced with RegCM and do not know if these modifications are easy, so this is more suggestion than a requirement. This suggestion touches also on the valid problem (already pointed out by referee #1) of removing dust by substituting desert cells by vegetated ones. Except the aforementioned albedo changes, there could be other unwanted interferences to the aerodynamic resistances and land-air interactions. I would think that the tweaking of dust emission through modifications of e.g. Eqs. 2, 3 in the dust module would be a much better technique.

We deeply appreciate this remark and thanks for the great suggestions. On one hand, the interaction of aerosols and EASM is very complicated, and there are many factors that can influence the monsoon. Contribution of dust aerosol to the monsoon variability may be only a small part. Perhaps the dust effect is significant only in the heavy dust years or strong monsoon years. Therefore, we focused on the heavy and light dust years. We agree that the meteorological fields could influence the results, especially in the light dust years with less dust aerosol. Thus, we compared the standard deviation of summer surface temperature and precipitation in CON with the signal induced by the dust effects (CON minus SEN) during the heavy dust years (Figure B2). The signal induced by the dust is much greater than the standard deviations. Therefore, the dust effects in our simulation is significant. We mentioned this in the revised version (Page 10, Line 17–22).

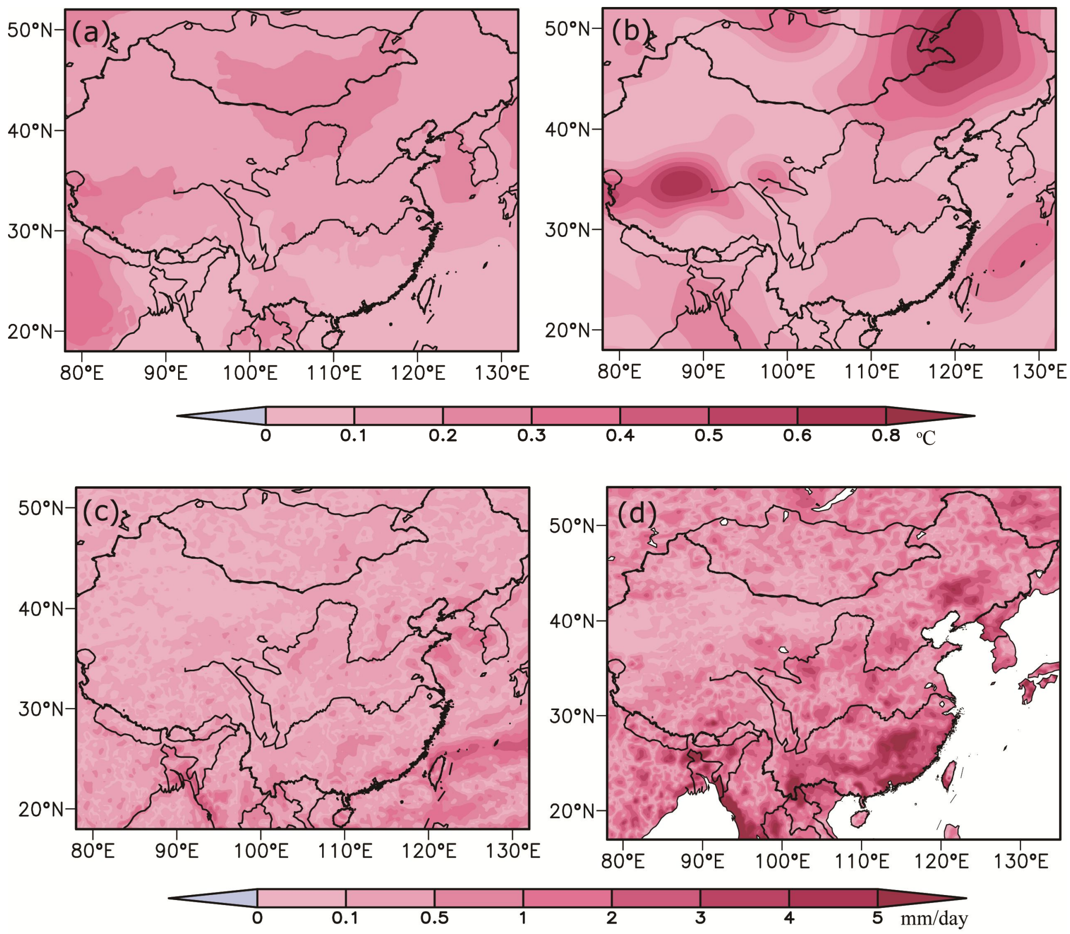

Figure B2 Standard deviation of summer surface temperature (a) and precipitation (c) in CON, and differences (absolute value) of summer surface temperature (b) and precipitation (d) between CON and SEN during the heavy dust years.

On the other hand, we agree that our method to turn off the dust emission in the TP is imperfect, and it will bring uncertainty to the results. Therefore, we carried out two new sensitive experiments to isolate effects of the changed land cover alone. Dust cycle in the two experiments is turned off, but the land cover is changed into one similar to the modification in SEN. The differences between them only include effects of the changed land cover. The results showed that the change (from desert to vegetated land) brought about 0.4 °C warming (Figure B3 and Figure B4), and this warming effect is weaker than the combined cooling effects (–0.6°C) induced by the dust aerosols and the changed land cover together. Besides, we have noted that Li and Xue (2010) demonstrated that the land cover change from vegetated land to bare ground (mainly desert) in the TP decreased the radiation absorbed by the surface and resulted in weaker surface thermal effects, which means that the effect of land cover change from desert to vegetated land may also cause a warming effects. This is opposite to the cooling induced by the dust aerosols over the TP in our simulation, and it may partly offset the dust aerosol induced cooling effects. However, our results show that the signal of cooling effect is not changed and it may be even underestimated in the heavy and light dust years. Therefore, actual cooling should be stronger than the simulated value. We admit that turning off the dust emission through modifications of Eqs. 2, 3 in the dust module would be a much better choice but it would be difficult to carry out given the short time frame. It is definitely worth trying in the future. We added discussion (Page 10, Line 22–34, and Page 11, Line 1–2) about uncertainty brought by the method used in this paper. The following texts were added in the discussion of the revised version.

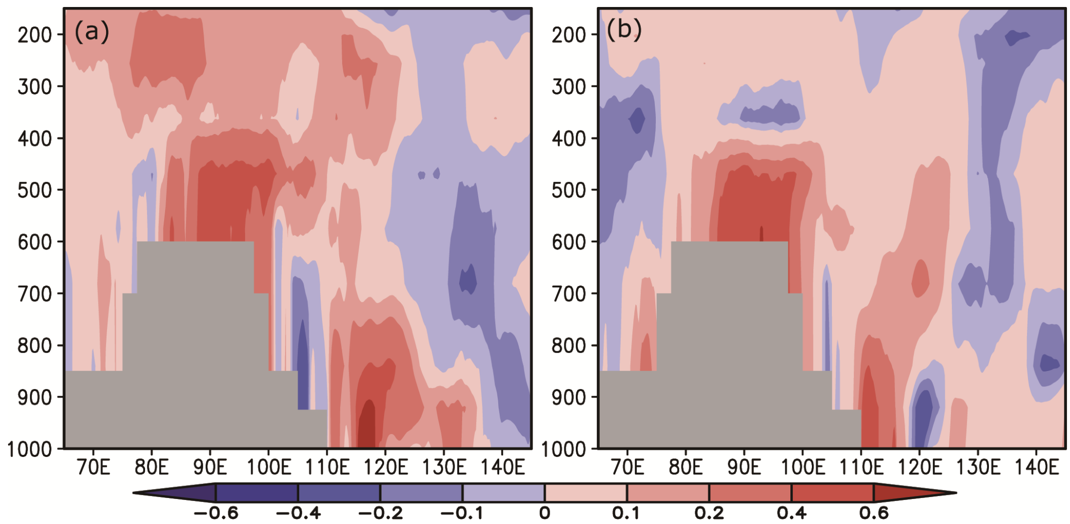

Figure B3 Longitudional-height cross-section of atmospheric temperature anomaly induced by the land cover changed averaged over 32–36°N in summer in heavy (a) and light (b) dust years.

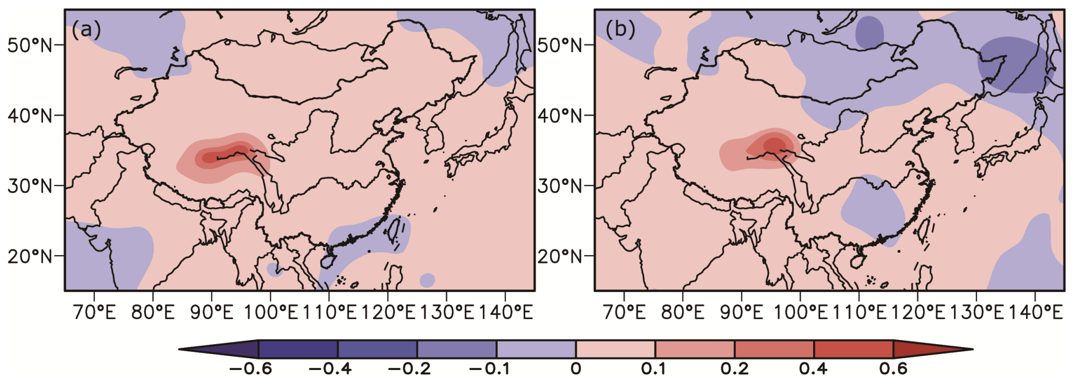

Figure B4 Summer temperature change induced by the land cover change in (a) heavy and (b) light dust years.

It is worth mentioning that the model's internal variability could influence the results; so we compared the standard deviation of summer surface temperature and precipitation in CON with the signal induced by the dust effects (CON minus SEN) during the heavy dust

years. The signal induced by the dust is much greater than the standard deviations (figures not shown). Therefore, the dust effect reported in our simulation is significant in the heavy dust years, but the cooling over central India in the light dust years may be caused by the model's internal variability. Besides, the results could also include the role of changed land cover in addition to the role of dust aerosols because turning off dust emission in the TP was through modifying underlying surface types. Hence, we carried out two addition sensitive experiments to isolate effects of the changed land cover alone. Dust cycle in the two experiments was turned off, but the land cover was changed into the one similar to the modification in SEN. The differences between them only included effects of the changed land cover. The results showed that the change (from desert to vegetated land) brought about 0.4 °C warming (figures not shown), and this warming effects is weaker than the combined cooling effects (–0.6°C) induced by the dust aerosols and the changed land cover together. Besides, it is interesting to note that Li and Xue (2010) demonstrated that land cover change from vegetated land to bare ground (mainly desert) in the TP decreased the radiation absorbed by the surface and resulted in weaker surface thermal effects, which mean that the effect of land cover change from desert to vegetated land may also cause warming effects. This is opposite to the cooling effect induced by the dust aerosols over the TP in our simulation, and the warming may partly offset the dust aerosol-induced cooling effect. However, our results showed that the signal of cooling effects was not changed, although it may even be underestimated. Therefore, actual cooling should be stronger than the simulated value. Hence, the reported dust effects also need be evaluated by using a refined way in the future.

Technical corrections:
My rather trivial corrections are listed below
p.1, l.20: Please use "stationary" instead of "stationery"

Taken care of. Thanks for the catch. (Page 1, Line 22)

p.1, l.26: Here "dust emission" is slightly better than "dust load".

Taken care of. Thanks for the catch. (Page 1, Line 27)

p.1, l.29: Please use "drivers of" instead of "drivers on".

Taken. Thanks for the catch. (Page 2, Line 1)

p.2, l.24: Please use "Gurbantunggut" instead of "Gubantunggut".

Taken care of. Thanks for the catch. (Page 2, Line 24)

p.2, l.27: Please use "elevated" instead of "elevate"

Taken care of. Thanks for the catch. (Page 2, Line 28)

p.4, Eq.4: χ and v are never defined

They are defined in the revised version as below (Page 4, Line 14).

χ is the dust mixing ratio, $\overline{V}$ is vector wind.

p.6, l.27: Please correct "respevtively" to "respectively"

Corrected. Thanks for the catch. (Page 6, Line 17)

p.9, l.22: Please use "spatiotemporal" instead of "spatiotemporal spatial"

Corrected. Thanks for the catch

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Direct radiative effects of dust aerosols emitted from the Tibetan Plateau on the East Asian summer monsoon – a regional climate model simulation

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Abstract. While dust aerosols emitted from major Asian sources such as Taklimakan and Gobi Deserts have been shown to have strong effect on Asian monsoon and climate, the role of dust emitted from Tibetan Plateau (TP) itself, where aerosols can directly interact with the TP "heat pump" because of their physical proximity both in location and elevation, has not been examined. This study uses the dust coupled RegCM4.1 regional climate model to simulate the spatiotemporal distribution of dust aerosols originating in the TP and their radiative effects on the East Asian summer monsoon (EASM) during both heavy and light dust years. Two 20-year simulations with and without the dust emission from TP showed that direct radiative cooling in the mid-troposphere induced by the TP locally produced dust aerosols resulted in an overall anticyclonic circulation anomaly in the low-troposphere centered over the TP region. The northeasterly anomaly in the EASM region reduces its strength considerably. The simulations found a significant negative correlation between the TP column dust load produced by local emissions and the corresponding anomaly in the EASM index (r=−0.41). The locally generated TP dust can cause surface cooling far downstream in eastern Mongolia and northeastern China through **stationary** Rossby wave propagation. Although dust from within TP (mainly Qaidam Basin) is a relatively small portion of total Asian aerosols, its impacts on Asian monsoon and climate seems disproportionately large, likely owning to its higher elevation within TP itself.

1 Introduction

Dust is one of the most important components of atmospheric aerosols. The main source of atmospheric dust is wind erosion in arid and semi-arid regions; it is estimated that the global atmospheric **dust emission** may be as high as 200–5000 Mt yr^{-1} (Goudie, 1983). Because of the large amount in the atmosphere, dust effects on the environment and the climate system have attracted much attention. The inhalation of dust aerosols can harm both human and animal health; it can also affect visibility and thus potentially increase the number of traffic accidents (Park and Kim, 2005). Dust aerosols also are important **drivers of** the global climate because of their direct radiative effects on the Earth–atmosphere radiation balance and temperature (Tegen and Lacis, 1996; Miller et al., 2004). They can alter the atmospheric hydrological cycle by acting as cloud condensation nuclei and thus can modulate both the regional and global precipitation (Rosenfeld, 2001). Satellite observations have shown that dust originating from the Taklimakan Desert can travel around the globe within two weeks and alter the interaction between the atmospheric CO_2 and the global climate by providing nutrients to and interacting with the marine ecosystem (Uno et al., 2009).

East Asia is an important source region for dust (Zhang et al., 1996) and is home to more than half of the world's population. The lives of people in East Asia are deeply affected by the East Asian summer monsoon (EASM) and the relationship between dust aerosols and the EASM is of great interest to the scientific community. Simulations have shown that dust aerosols not only weaken the EASM (Sun et al., 2012; Guo and Yin, 2015), but can also reduce the atmospheric heat source over the Tibetan Plateau (TP) and delay the onset of the EASM (Sun et al., 2016). Aerosols, including dust aerosols, have been shown to affect the intensity of the EASM (Li et al., 2016) and variations in the EASM can modulate the spatiotemporal distribution of dust aerosols in East Asia. A recent modeling study by Lou et al. (2016) indicated that there was a negative correlation between the spring dust loading in eastern China and the East Asian monsoon.

Dust aerosols in East Asia are mainly derived from arid and semi-arid areas, including the Taklimakan Desert and the Gobi Desert. However, some studies have indicated that the TP itself may also be an important source region for dust (Zhang et al., 1996; Fang et al., 1999) and that the region is more conducive to the atmospheric transportation of dust due to its high altitude and it can interact directly with the TP thermal pump (Wu et al., 2012). However, the source and spatiotemporal distribution of dust aerosols over the TP have not been established yet. At present, there are three viewpoints about the source of dust aerosols over the TP. First, an investigation by Fang et al. (1995, 1999) showed that there exists 2047.41×10^4 km^2 of desert land over the TP, suggesting that the TP may be a potential source for dust. A numerical simulation by Chen et al. (2013) showed that dust aerosols were produced by local emissions over the TP in spring and winter. Second, satellite observations have shown that the aerosols over the TP are dominated by dust in spring and summer and that the dust aerosols were probably derived from the Taklimakan Desert and the **Gurbantunggut** Desert to the north of the TP (Huang et al., 2007; Jia et al., 2015). Third, some studies have indicated that the dust emitted from the south of the TP, such as from the Great Indian Desert, can also be transported over the Himalaya (Lau et al., 2006, 2010).

As a massive, **elevated** heat source, the TP can directly heat the upper troposphere. The heating anomaly over the TP has a great impact on the EASM (Yanai and Wu, 2006; Duan et al., 2012). Studies have shown that dust aerosols over the TP can alter the local atmospheric radiation balance, affecting both the heat source over the TP and the Asian monsoon (Lau et al., 2006, 2010; Chen et al., 2013; Sun et al., 2016). However, most previous simulation studies have focused on dust aerosols originating from the Taklimakan and Gobi deserts (Zhao et al., 2006; Wang et al., 2008; Huang et al., 2009; Sun et al., 2012) and there have been few investigations of the impact of dust aerosols emitted by the TP on the East

Asian climate. The work reported here used the RegCM4.1 model to simulate climatic effects of distribution of dust aerosols surrounding the TP by performing numerical experiments with and without the emission of dust over the TP.

2 Numerical model and experiment design

2.1 RegCM4.1 model

We used the RegCM4.1 model (Regional Climate Model version 4.1), which is developed and supported by the National Center for Atmospheric Research (NCAR) and the International Center for Theoretical Physics. The model has been widely used for more than 20 years in studies of regional climatic and environmental change, especially in the simulation of the effect of aerosols on climate (Qian et al., 2003; Solomon et al., 2008; Zhang et al., 2009; **Zanis et al., 2012; Ji et al., 2011, 2015; Das et al., 2015a, 2016; Mbienda et al., 2017**).

The dynamic framework of RegCM4.1 core is based on the hydrostatic core of the mesoscale model MM5 (Grell et al., 1994). The radiation scheme in RegCM4.1 is the CCM3 radiation transfer process (Kiehl et al., 1996). RegCM4.1 has two land surface process schemes: (1) the biosphere atmosphere transfer scheme (BATS1e) (Dickinson et al., 1993); and (2) the Common Land Surface process module (CLM3.5) (Oleson et al., 2008). The dust cycle can only be diagnosed when BATS1e is used. The planetary boundary layer parameterization in RegCM4.1 follows the scheme of Holtslag et al. (1990) and there are three cumulus convection parameterization schemes, including Grell (Grell et al., 1993), Kuo (Anthes, 1977) and MIT-Emmanuel (Emanuel, 1991). The dust module coupled in RegCM4.1 is based on the dust emission model (DPM) of Marticorena et al. (1995) and Alfaro and Gomes (2001). It considers dust emission, dry/wet deposition and the diagnosis of the optical and radiation characteristics of dust (including long- and short-wave radiation) (Zakey et al., 2006; Zhang et al., 2009). RegCM4.1 uses the δ-Eddington approximation for radiative flux calculations and the calculation of dust short-wave radiation uses an asymmetry factor, single scattering albedo and mass extinction coefficient based on the Mie theory. The dust long-wave radiation is accounted for by introducing the dust emissivity given by Kiehl et al. (1996).

The coupled dust module has been described in detail in previous articles (Zakey et al., 2006; Zhang et al., 2009); so only a brief introduction is given here. There are four steps in dust parameterization. First, each model grid cell is classified as either desert or non-desert according to its soil properties (such as texture, soil type, particle size and composition) based on the United States Department of Agriculture textural classification. Second, dust emission is assumed to be a function of friction velocity (u^*); dust aerosols is lifted off the ground once u^* exceeds a threshold value ($u_t^*(D_p)$).

$$u_t^*(D_p) = u_{ts}^*(D_p) \cdot f_{eff} \cdot f_w \,, \qquad (1)$$

there $u_{ts}^*(D_p)$ depends on soil particle size(D_p), f_{eff} and f_w are the correction terms for nonerodible surface roughness elements (Marticorena and Bergametti, 1995) and soil moisture content (Fecan et al., 1999), respectively.

Third, the horizontal mass flux is treated as a function of the frictional velocity and is given by:

$$dH_F(D_p) = E \cdot \frac{\rho_a}{g} \cdot u^{*3} \cdot (1 + R(D_p)) \cdot (1 - R^2(D_p)) \cdot dS_{rel}(D_p), \quad (2)$$

where E, ρ_a and g are the ratio of the erodible to total surface areas, the surface air density and the gravitational acceleration, respectively. $R(D_p)$ is the ratio of the threshold frictional velocity in equation (1) to the frictional velocity u^* calculated within each grid cell from the model prognostic surface wind and surface roughness height. $dS_{rel}(D_p)$ is the relative surface area of a soil aggregate of diameter D_p to the total surface area of soil aggregates. The vertical flux corresponding to each emission mode is calculated by:

$$F_{dust,i}(D_p) = (\frac{\pi}{6}) \cdot \rho_p \cdot D_i^3 \cdot N_i, \quad (3)$$

where ρ_p is the aggregate density (2.65 g cm^{-3}), D_i is the median diameter and N_i is a function of the kinetic energy flux.

The dust particles are divided into four size bins (or modes): fine (0.01–1.0 μm), accumulation (1.0–2.5 μm), coarse (2.5–5 μm) and giant (5.0–20.0 μm). The dust transport, deposition and removal processes are given by Qian et al. (2001) and Qian and Giorgi (1999):

$$\frac{\partial \chi^i}{\partial t} = -\overline{V} \cdot \nabla \chi^i + F_H^i + F_V^i + F_C^i + S^i - R_{Wls}^i - R_{Wc}^i - D_d^i, \quad (4)$$

where **χ is the dust mixing ratio, $\overline{V}$ is vector wind, and** $-\overline{V} \cdot \nabla \chi^i$ is the advection, F_H^i is the horizontal turbulent diffusion, F_V^i is the vertical turbulent diffusion and F_C^i is the convective transport. R_{Wls}^i and R_{Wc}^i are the wet removal terms, represented by large-scale and convective precipitation. D_d^i is the dry deposition, represented by assuming fixed depositional velocities over both land and water.

2.2 Experimental design and observational data

Two numerical experiments were designed; both integrate for 20 years (excluding first two years of spin-up) using the dust-coupled RegCM4.1. The first experiment was a control experiment (CON) that used the default land use types from the Global Land Cover Characterization dataset (Loveland et al., 2000), meaning that dust emitting sources both within and outside the TP. The second experiment was a sensitivity experiment (SEN) where we turned off the dust emission in the northern and northeastern TP (the deserts inside the black outline of the TP contour in Fig. 1b). To eliminate the dust emission in the TP, we replaced the land cover types of these areas with the nearby vegetated types, in a manner similar to the method of Liu et al., (2015a). All the other conditions in the sensitivity experiment were the same as in the control experiment. **In order to isolate the effect of dust aerosols, only dust aerosols are included in our simulations, without considering other aerosols (such as anthropogenic or marine aerosols).**

The initial and boundary conditions were taken from the NCAR/NCEP re-analysis dataset (Kalnay et al., 1996). The sea surface temperature used the National Oceanic and Atmospheric Administration sea surface temperature dataset (Reynolds et al., 2002). The topography of the TP is very complex, necessitating high spatial resolution (<60 km) to resolve localized precipitation (Gao et al., 2006). The horizontal resolution in RegCM4.1 runs was therefore set to 40 km. The simulation domain is shown in Fig. 1a and the model domain centered was at (32°N, 105°E), with 240 grid cells in the west–east direction and 160 grid cells in the north–south direction, respectively. The model was run in the standard configuration of 18 vertical σ layers with the model top at 10 hPa. The integration duration for both experiments was from January 1, 1988 to December 31, 2009. The first two years were treated as the model spin-up time and only the results from the last 20 years were analyzed.

Five main types of observations were used to evaluate the simulated results of CON: (1) the monthly mean surface air temperature and precipitation, with a high resolution of 0.5×0.5°, provided by the Climate Research Unit (CRU) of the University of East Anglia (Mitchell and Jones, 2005), which was used to evaluate the simulated surface temperature and precipitation in CON; (2) the NCEP–DOE re-analysis wind field (2.5×2.5°) at 850 hPa, which was used to compare the simulated atmospheric circulation; **(3) level-3 monthly mean AOD data during 2000 to 2009 obtained from the Multiangle Imaging Spectroradiometer (MISR) onboard NASA's Earth Observation System Terra satellite (http://www-misr.jpl.nasa.gov/). Since MODIS AOD has a large portion of missing data in Northwest China, MISR was used to evaluate the simulated dust AOD in CON. The effectiveness of the MISR data was investigated by Martonchik et al. (1998, 2004) and Bibi et al. (2015). (4) level-3 monthly mean pure dust AOD data under cloud free scenes (2×5°) from 2007 to 2009 obtained from Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) (Winker et al., 2013), which was also used to evaluated the simulated dust AOD in CON. The most recent version of the L3 product included averaging of individual types (Liu et al., 2008; Amiridis et al., 2013; Marinou et al., 2016)**; and (5) the AOD observed *in situ* by the Aerosol Robotic Network (AERONET), which was used to evaluate the simulated dust seasonal and interannual variation in CON.

3 Results of simulations

In this section we will first evaluate the CON simulation using the observed data described in the previous section. Then the results from CON and SEN experiments will be compared to determine the roles of dust aerosols generated from the TP play in the thermodynamic fields and circulations including the EASM.

3.1 Validation

3.1.1 Basic model climatology

The simulated climatology can influence the distribution of dust aerosols and their climatic effects, so CON was used to analyze the surface temperature, precipitation and atmospheric circulation at 850 hPa. The CON-simulated and CRU-observed 20-year average summer surface temperatures in East Asia are presented respectively in Figs 2a and 2b. The CRU observed temperature is >25°C in southern China, NW China and northern India, and <7°C over the TP. The observed north–south gradient and location of the maximum and minimum centers were captured well by RegCM4.1. The model captured the major distribution patterns of precipitation, including the reasonable SE–NW gradient and the maximum centers in southern China, the Himalaya and Indian Peninsula, with a 2–4 mmday^{-1} negative bias in the Korean Peninsula and south Japan and a 2–4 mmday^{-1} positive bias in the Tianshan Mountains (Figs. 2c, 2d). These simulated deviations are likely related to the cumulus convective scheme in the model (Zhang et al., 2008; Wang and Yu, 2011). RegCM4.1 captured the major characteristics of the circulations in East Asia, where southwesterlies dominate to the south side of the TP, and the location of the Indian Low is consistent with the NCEP–DOE observations (Figs. 2e, 2f).

3.1.2 Dust AOD comparison between simulated and MISR observed

Satellite and *in situ* observations include all types of aerosols, such as black aerosols, SO_2 and organic carbon; observed data for dust AOD alone are scarce. Therefore we used the MISR AOD data, as in most previous studies (Zakey et al., 2006; Zhang et al., 2009), to evaluate the spatiotemporal distribution of the dust AOD simulated by the model. Both the simulation and observations showed that the dust AOD over the TP and its surrounding areas was higher in spring and summer (Figs. 3a and 3c) and lower in autumn and winter (Figs. 3e and 3g). There were three maximum centers (maximum value >0.6) of dust AOD in spring and summer, located in the Taklimakan Desert, the Gobi Desert and the Great Indian Desert, **respectively**. The dust AOD over the Qaidam Basin in the NE of the TP was also >0.5 and the dust AOD over the northern TP, adjacent to the southern Taklimakan Desert, was between 0.3 and 0.5. The simulated dust AOD in these regions was reduced in autumn and winter (Figs. 3e and 3g). The MISR-observed AOD was largely consistent with the model results for the Taklimakan Desert, the Gobi Desert and the Qaidam Basin, but were relatively low in the Great Indian Desert in summer. The large value of the MISR AOD in the Sichuan Basin to the east of the TP was due to industrial emissions, which were not incorporated into our model simulation.

3.1.3 Dust AOD comparison between simulated and AERONET-observed

Figure 4 compares the *in situ* observed monthly mean AOD from AERONET and that simulated by RegCM4.1 at Dalanzadgad (43.6°N, 104.4°E). This is the only available AERONET site in the vicinity of the dust sources with continuous records for >10 years. The model captured the seasonal and interannual variations of AOD well, including the year with extremely high levels of dust. Observations over the TP are scarce and we could only find a site with continuous aerosol records from AERONET at Nam Co (30.77°N, 90.96°E). The seasonal variation of AOD at this site is well captured. Both the simulation and the observations showed that the dust AOD increases in spring at Nam Co (Fig. 5).

3.1.4 Simulated and CALIPSO-observed dust AOD comparison

While MISR and AERONET data contain all types aerosols including those anthropogenic ones, the CALIPSO observation sole devotes to dust aerosols. Figure 6 shows that the simulated seasonal variation, center positions and magnitude of dust AOD are very consistent with those observed by CALIPSO during day and night. Both simulations and observations not only showed that dust AOD increased in spring and summer and decreased in autumn and winter, but also captured three maximum centers of dust AOD in Taklimakan, the Great Indian Desert and Qaidam Basin located in the northern TP in spring. The simulated center values were still high in summer. Besides, it is very interesting to note that the observed dust AOD in the Qaidam Basin is higher at night than that during daytime (Figure 6b and 6c), which implies that dust activities in the TP may be more prominent at night. This unusual feature could cause dust radiative effects in the TP is very different from those of in other locations, which is further discussed in the subsequent section of 3.4.1

3.2 Relationship between the EASM and dust loading over the TP

To study the relationship between dust aerosols and the EASM, we used the average summer meridional wind at 850 hPa over eastern China (20–45°N, 105–122.5°E) as an EASM index, following Xie et al. (2016). **This index measures the intensity of the southerly wind to the east of TP in lower troposphere above East Asia.** It has been widely used to examine both modern and paleo-changes in the East Asian monsoon (Wang et al., 2008; Jiang and Lang, 2010). We found that the simulated difference in the EASM index (CON−SEN) and the difference in the model-simulated column dust load averaged over the TP are highly anticorrelated with a correlation coefficient $R=-0.41$ (Fig. 7). The dust aerosol increases and decreases over the TP as the index weakens and enhances **respectively**. Lou et al. (2016) also demonstrated a clear negative correlation between the EASM and the dust concentration over eastern China in spring. Based on the variation in the column dust load shown in Fig. 7, we chose 1994 and 2009 as heavy dust years and 2003 and 2007 as light dust years and then contrasted the dust distribution over the TP and its effects on the summer climate in the heavy/light dust years.

3.3 Dust aerosol distribution in heavy/light dust years

In the heavy dust years, the difference in the column dust load over the TP was greater than that in the light dust years, as expected. Two centers of maximum column dust load existed over the TP in the heavy dust years (Fig. 8a). One was located in the Qaidam Basin and the other in the NW of the TP. The maximum values at both centers were >70 mg m^{-2}. However, the difference in the column dust load over the NW of the TP in the light dust years was much lower than in the heavy dust years and the central value was <25 mg m^{-2}. From the vertical profiles of the dust load (Figs 8c and 8d), we can see that the dust mixing ratio was higher in heavy dust years in the western TP with a central value >5 μg kg^{-1}. The mixing ratio was lower in the western TP in the light dust years.

3.4 East Asian climate anomalies in heavy/light dust years induced by dust aerosols

3.4.1 Temperature anomaly

Both the atmospheric heating rate and the atmospheric temperature over the TP decreased in heavy/light dust years (Fig. 9). During the heavy dust years, the dust aerosol resulted in two cooling anomaly centers, one in the lower troposphere (600–400 hPa) over TP core and the other in the upper troposphere (*c.* 300 hPa) of the western TP, with a cooling of >0.4 K day^{-1} due to the large dust load. These cooling anomalies resulted in a low temperature center at 500 hPa over the western TP, with its central value reduced by >0.6°C. The dust aerosol over the TP in the light dust years was much less than in the heavy dust years (Fig. 8a and 8c), so the cooling effect in the light dust years was weaker than in the heavy dust years. The warming to the east of the TP (110–120°E) (Fig. 9c and 9d) is caused by diabatic heating induced by the enhancement of descending motion. **The enhanced descending motion was induced by the LW cooling effects of dust over the TP (figures not shown)**.

The surface temperature over the TP decreased in both the heavy and light dust years (Fig. 10). In the heavy dust years, the surface temperature decreased by 0.6°C over the TP and the sea–land thermal contrast was reduced. It is worth noting that the effects of TP aerosols on surface temperature were not limited to local or surrounding regions. In fact, the largest impact was in NE China, more than two thousands km downstream (Fig. 10a). The remote cooling is likely contributable to a cold air advection stemmed from the upstream TP aerosols to be discussed in next subsection (Fig. 11). **Similar phenomena were reported earlier from Europe for anthropogenic aerosols (Zanis 2009; Zanis et al., 2012) and from South Asia for natural aerosols (Das et al., 2015a).**

3.4.2 Circulation

The overall effects of TP aerosols cool the troposphere surrounding the TP (Fig. 10a) and thus the land–sea thermal contrast was reduced by the dust aerosols over the TP. The atmospheric circulation anomaly induced by the dust aerosols emitted over the TP in heavy dust years shows an overall gigantic anticyclonic circulation centered over the TP **with a positive anomaly (>10m) in geopotential height** (Fig. 11a). The northeasterlies that run against the southwesterly monsoon is especially strong over the EASM region, which indicates that the EASM was weakened greatly. The anomaly still existed in the light dust years, but its intensity was much weaker than in the heavy dust years (Fig. 11b).

3.4.3 Precipitation

Figure 12 shows the simulated change in summer precipitation in **East Asia** induced by dust emitted over the TP in heavy and light dust years. The precipitation decreased **in both the southern and the northern monsoon regions** in summer in the heavy dust years as a result of weakening of the EASM (Fig. 11), **and the reduction in the southern monsoon region is greater than that in the northern monsoon region**. The dust aerosols also reduced precipitation in the two monsoon regions in the light dust years. The simulated suppressive effects of the dust aerosols were consistent with previously reported modeling results (Sun et al., 2012; Guo et al., 2015). **Besides, precipitation in the heavy dust years reduced more than that in the light dust years in the TP, which may be suppressed by the enhancement of descending motion induced by the strong cooling effects of dust aerosol over the TP.**

4 Discussion

Previous research has shown that dust emitted from Asian deserts can weaken the EASM (Sun et al., 2012; Guo et al., 2015; Li et al., 2016), although the details of weakening mechanisms are still unclear. It has been suggested by some authors that the weakening of the EASM is a result of the reduction in the thermal contrast between the land and the sea induced by dust aerosols (Guo and Yin, 2015; Li et al., 2016). However, the modeling result of Sun et al. (2012) showed that the EASM is reduced by the large-scale atmospheric disturbances (cyclone–anticyclone–cyclone Rossby wave train) generated by the radiative cooling of dust aerosols. In the work reported here, we considered the effects of dust aerosols emitted only within the TP itself on regional climate and found that they can also reduce the EASM significantly by weakening the heat source ("pump") over the TP and thus reduce the land–sea thermal contrast. The locally generated TP dust can cause surface cooling far downstream in eastern Mongolia and northeastern China through stationary Rossby wave propagation. The dust-loading over the TP include both local emissions and external sources; our sensitivity simulations showed there was a negative correlation between the EASM and dust aerosols emitted from the TP locally.

The spring dust aerosols from the TP have a close relation with EASM. Although the cause-effect relationship is not immediately clear, the following processes are proposed as a possible mechanism of this relation based on the results in our simulation (Fig. 13). Firstly, increasing (decreasing) in dust aerosol over the TP in the heavy (light) dust years in spring can weaken (enhance) the TP heat source and thus reduce (increase) precipitation over the TP. Reduction (increase) in precipitation over the TP can also further enhance (diminish) dust emission over the TP. Secondly, the weakened (enhanced) TP heat source can persist from spring to summer and shrink (expand) the land-sea thermal contrast and thus weaken (enhance) the EASM. Therefore, the change of dust over the TP has an anti-correlation with the variation of EASM circulation intensity. Thirdly, weakened (enhanced) monsoon circulation can reduce (increase) precipitation in East Asia. As a result, the precipitation variation of the TP presents a positive-correlation with that of EASM.

It is worth noting that Sun and Liu (2016) demonstrated that dust emitted from Taklimakan and Gobi Deserts weakens the Asian monsoon through large-scale atmospheric disturbances by 2 m s^{-1} of wind at 700 hPa. This magnitude of reduction in the wind seems small compared to the values in the present study even though the emission source extent in the previous is larger. We think both high-altitude source like the TP and low-altitude sources such as Taklimakan and Gobi Desert can weaken the EASM, but the mechanism could differ. The dust emitted from low altitude source (mainly Taklimakan and Gobi Desert) reduces the EASM mainly by the large-scale atmospheric disturbances, while the dust emitted from high-altitude source weakens the EASM by the reduction in the TP heating and in thermal contrast in the middle troposphere between the land and sea. The column dust load induced by local emissions from the TP in heavy dust years accounted for 20% ((CON−SEN)/CON) of the total loading over the TP, its impacts on Asian monsoon and climate seems more important than the low altitude sources such as Taklimakan and Gobi Desert in East Asia. This disproportionately large impact from TP locally emitted dust is likely due to its higher elevation within TP itself so that the dust-induced cooling can more effectively weaken the TP's acting as a heat pump for the Asian monsoon. Further studies on this is rightly warranted.

Dust direct radiative effects is reported predominantly positive (warming) over land areas in most of previous researches (Zhang et al., 2013; Chen et al., 2013), but it can become negative under specific situations like a large zenith angle (Quijano et al., 2000). It is interesting to note that it can also become negative when dust activities are mainly vigorous at night over the TP. The observed dust AOD at night is much higher than those of during daytime over the TP (Figure 6b and 6c); so the SW warming effects is quite weaker than the LW cooling effects (figures not shown). The simulations of Chen et al. (2013) included strong absorbing aerosols such as black carbon and included contribution of other dust sources including such as the Taklimakan and the Gurbantunggut Desert to the north of the TP and that from the Great Indian Desert to the south of the TP; thus, the warming effects of dust aerosol reported in their study is broader. In contrast, the cooling effects in our study is mainly caused by the dust aerosol emitted from the TP at night.

It is worth mentioning that the model's internal variability could influence the results; so we compared the standard deviation of summer surface temperature and precipitation in CON with the signal induced by the dust effects (CON minus SEN) during the heavy dust years. The signal induced by the dust is much greater than the standard deviations (figures not shown). Therefore, the dust effect reported in our simulation is significant in the heavy dust years, but the cooling over central India in the light dust years may be caused by the model's internal variability. Besides, the results could also include the role of changed land cover in addition to the role of dust aerosols because turning off dust emission in the TP was through modifying underlying surface types. Hence, we carried out two additional sensitive experiments to isolate effects of the changed land cover alone. The dust cycle in the two experiments was turned off, but the land cover was changed to one similar to the modification in SEN. The differences between them only included effects of the changed land cover. The results showed that the change (from desert to vegetated land) brought about 0.4 °C warming (figures not shown), and this warming effects is weaker than the combined cooling effects (–0.6°C) induced by the dust aerosol and the changed land cover together. Besides, it is interesting to note that Li and Xue (2010) demonstrated that the land cover change from vegetated land to bare ground (mainly desert) in the TP decreased the radiation absorbed by the surface and resulted in weaker surface thermal effects, which means that the effect of land cover change from desert to vegetated land may also cause warming effects. This is opposite to the cooling effect induced by the dust aerosols over the TP in our simulation, and the warming may partly offset the dust aerosols-induced cooling effect. However, our results showed that the signal of cooling effects was not changed, although it may even be

underestimated. Therefore, actual cooling should be stronger than the simulated value. Hence, the reported dust effects also need be evaluated by using a refined way in the future.

It is very beneficial to study the impact of aerosols on climate using a RCM instead of a coarse-resolution global climate model (GCM). The GCMs tend to systematically underestimated dust aerosol concentration, presumably due to their lower spatial resolution (Tegen et al., 2002; Zhang et al., 2009). The RCM simulated dust concentration, on the other hand, was closer to the observed magnitude compared with the results of global models (Sun et al., 2012). For example, previous studies showed that high-resolution RCMs had better capabilities than GCMs in simulating the effect of aerosols on Asian monsoon (Zhou and Yu, 2006; Gao et al., 2008; Ji et al., 2011). A high-resolution regional model is especially needed to capture subtle characteristics in the areas of complex terrain (Ji et al., 2011). However, limited area RCM naturally cannot fully account for external forcing remote from the domain of interest although the lateral boundary conditions allow large-scale features to propagate into the domain. Our domain size (9600 km × 640 km) is reasonably large enough so that the weather and climate systems can have adequate spatial extent to develop within the domain, as attested by reasonably validation of wind pattern, temperature field, and precipitation (Section 3.1). Cautions should be exercised, however, that results from regional simulations could be somewhat domain-size dependent quantitatively although main results should not be affected.

Only direct radiative effects of dust were included in our model and future studies should include both direct and indirect effects. The simulated effects of dust aerosols on climate were highly sensitive to the physical characteristics of the dust aerosols, such as the single scattering albedo (Huang et al., 2014; Colarco et al., 2014; Das et al., 2015b). Therefore our results also need to be validated by sensitivity experiments using aerosols with different properties. **A recent study by Tsikerdekis et al. (2017) demonstrated that simulated dust load and induced radiation change are sensitive to the dust particle size division in the model; so further sensitivity experiments using more dust size bins would be worthwhile.** In addition, many other factors can also affect the EASM, including the El Niño Southern Oscillation (Zhao et al., 2012; Liu et al., 2015b), the North Atlantic Oscillation (Wu et al., 2009) and heat sources over the TP (Yanai and Wu, 2006; Duan et al., 2012). A recent numerical simulation by Wang et al. (2017) showed that aerosol emissions from outside East Asian play an important part in weakening the circulation of the EASM.

5 Conclusions

We conducted two numerical experiments to quantify the effects of dust aerosols emitted over the TP on the EASM in heavy/light dust years using a high-resolution regional climate model. Satellite and *in situ* observations were used to evaluate the simulated spatial distribution of dust aerosols and their seasonal and interannual variations. We analyzed the change in dust aerosols induced by emissions over the TP and their radiative effects on the EASM and summer precipitation in heavy/light dust years. We also studied the effects of dust aerosols on the onset of the EASM.

The spatiotemporal distribution of the dust AOD and their seasonal and interannual variation were captured well by the RegCM4.1 model compared with the MISR AOD and *in situ* observations from AERONET. Both the simulated and observed AOD were higher in spring/summer and lower in autumn/winter. The simulated dust AOD was higher in the Taklimakan Desert, the Gobi Desert and the Great Indian Desert, with peak values >0.6. The simulated dust AOD in the Qaidam Basin and the northern TP were also higher. The seasonal variation in the dust AOD at Nam Co was captured well by RegCM4.1 relative to the observed aerosol AOD.

Comparative analyses of the two simulations indicated that the dust aerosols generated over the TP had a profound influence on the EASM. The difference in the EASM index and column dust load between CON and SEN experiments are negatively correlated (r=−0.41). The index also weakened (enhanced) as the imported-local combined dust aerosol increased (decreased) over the TP. The net atmospheric heating rate was negative over the TP in heavy dust years as a result of the radiative cooling effects of the dust aerosols, leading to a 0.6°C cooling in the surface and atmospheric temperatures. The land–sea thermal contrast and EASM were therefore both weakened, causing a 12% reduction in precipitation in the southern monsoon region. The dust load over the TP in the light dust years was much less than in the heavy dust years, implying large interannual variability. The dust aerosols produced over the TP can delay the onset of the EASM by one pentad in both the northern and southern monsoon regions of China.

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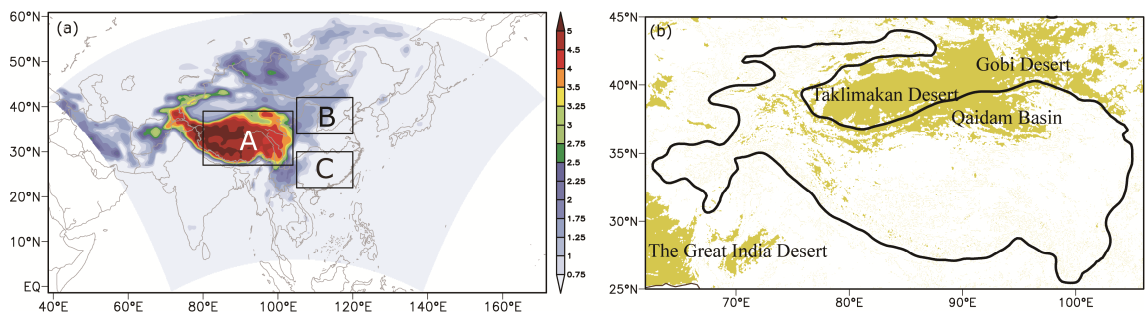

Figure 1: (a) Model domain and topography (units: km) and (b) dust source regions (yellow area) over the Tibetan Plateau and surrounding areas. Rectangles in (a) indicate areas Tibetan Plateau (A, 27–39°N, 80–105°E), north EASM region (B, 34–42°N, 105–120°E), and south EASM region (C, 22–30°E, 105–120°N).

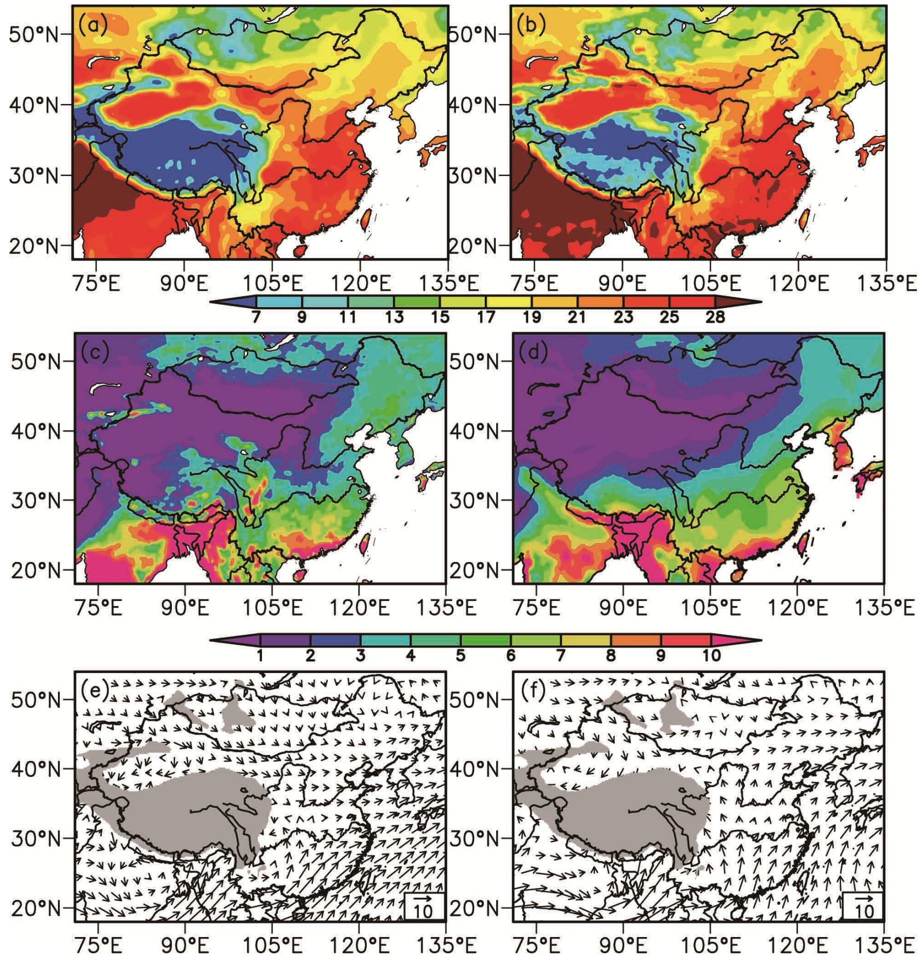

Figure 2: Spatial distribution of (a, b) **summer** surface air temperature (°C) and (c, d) **summer** precipitation (mm day^{-1}) simulated in the control experiment (left) and the CRU observations (right) for 1990–2009. The bottom two panels are wind vectors at 850 hPa simulated in (e) the control experiment and (f) the NCEP–DOE re-analysis during the summer monsoon season (June–August) averaged for 1990–2009.

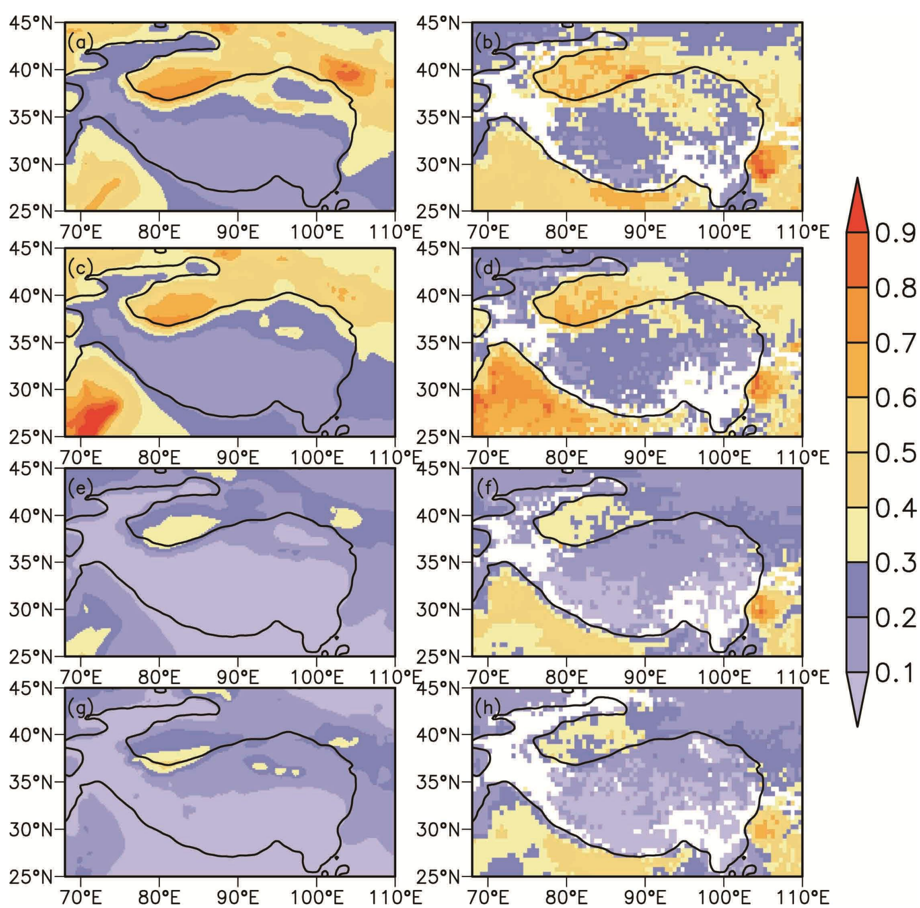

Figure 3: Spatial distribution of the dust AOD simulated by the control experiment (left panels) and the total aerosol optical depth observed by MISR at 550 nm (right panels) averaged in (a, b) spring, (c, d) summer, (e, f) autumn and (g, h) winter during the time period 2000–2009.

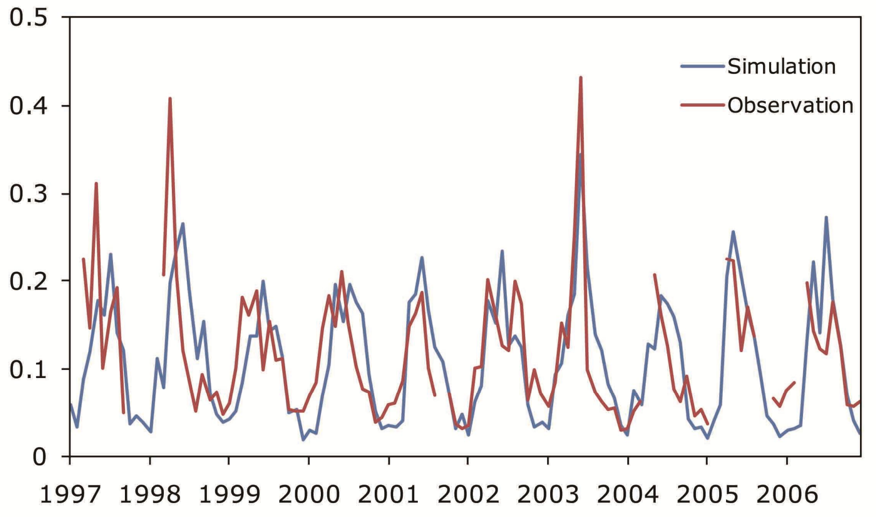

Figure 4: Comparison between the simulated variation of the monthly mean dust AOD in the control experiment and the AERONET-observed variation of the monthly mean aerosol AOD (500 nm) at Dalanzadgad from 1997 to 2006 (***R*****=0.66**).

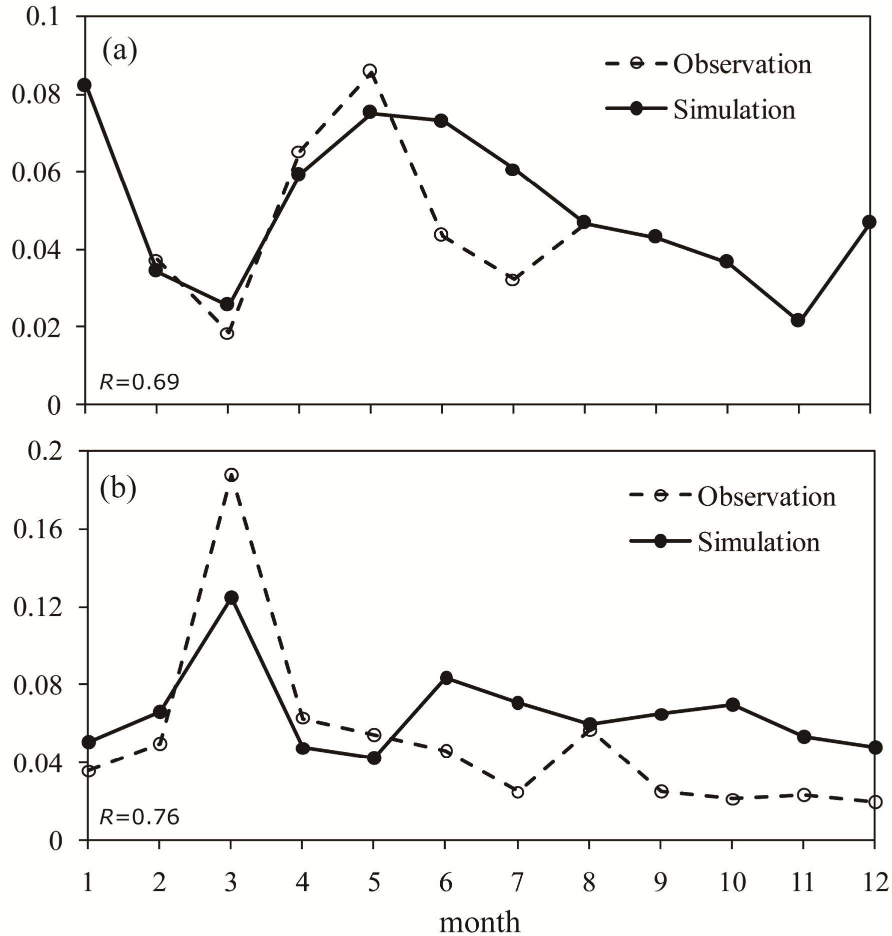

Figure 5: Comparison between the AERONET-observed monthly mean aerosol AOD (500 nm) at Nam Co and simulated by the control experiment at the grid near Nam Co in (a) 2007 and (b) 2009.

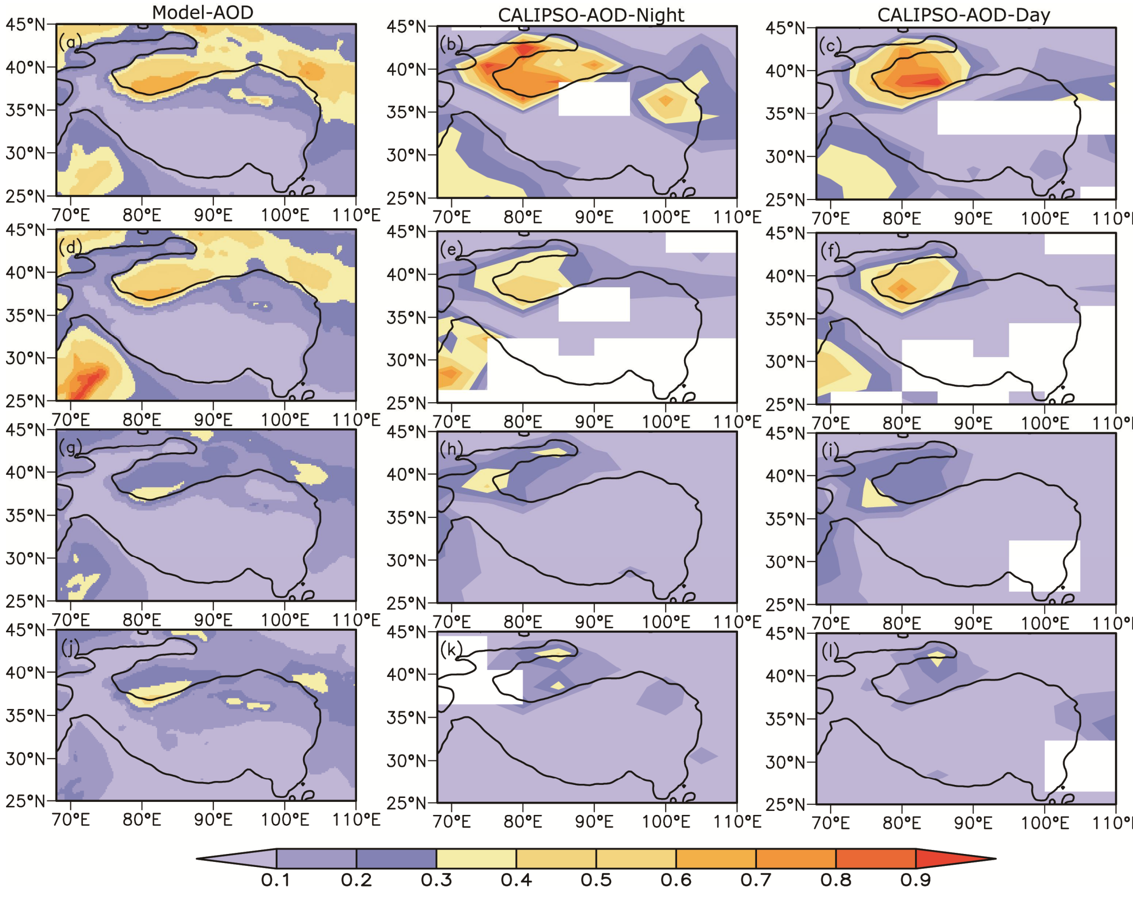

Figure 6: Spatial distribution of the dust AOD simulated by the control experiment (left panels) and observed by CALIPSO at night (middle) and during daytime (right panels) averaged in (a, b, c) spring, (d, e, f) summer, (g, h, i) autumn and (j, k, l) winter during the time period 2007–2009.

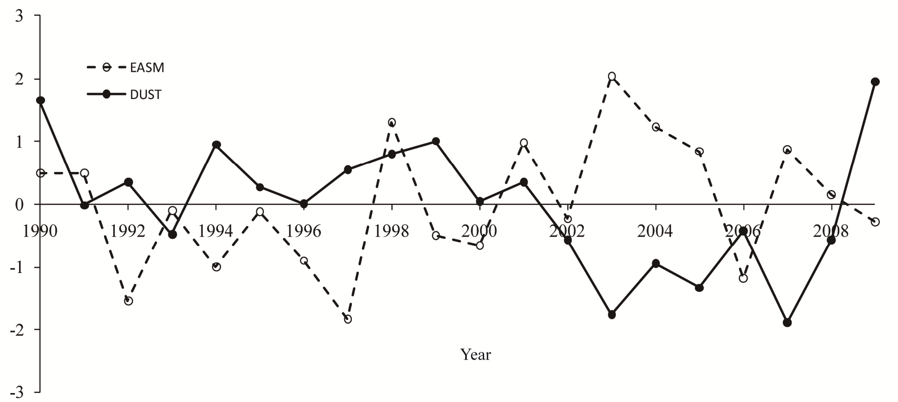

Figure 7: Difference (CON minus SEN) in the normalized regional mean dust column load averaged over the Tibetan Plateau (27–39°N, 80–105°E) and in the EASM index for summer during the period 1990–2009 (*R*=−0.41).

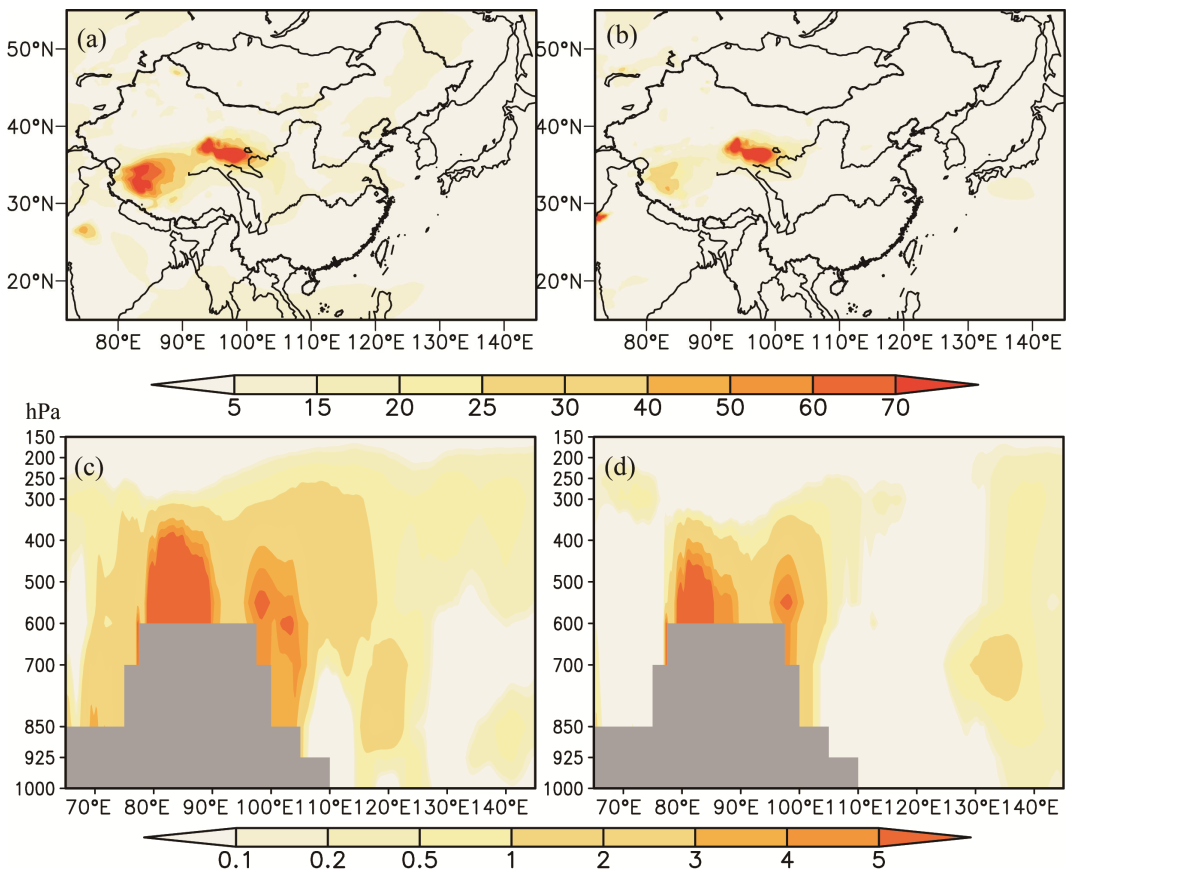

Figure 8: (a, b) Simulated differences (CON minus SEN) in the horizontal distribution of the column dust load (mg m^{-2}) and (c, d) the longitude–height cross-section (averaged over 32–36°N) of the dust mixing ratio (μg kg^{-1}) for summer in heavy (left) and light dust years (right).

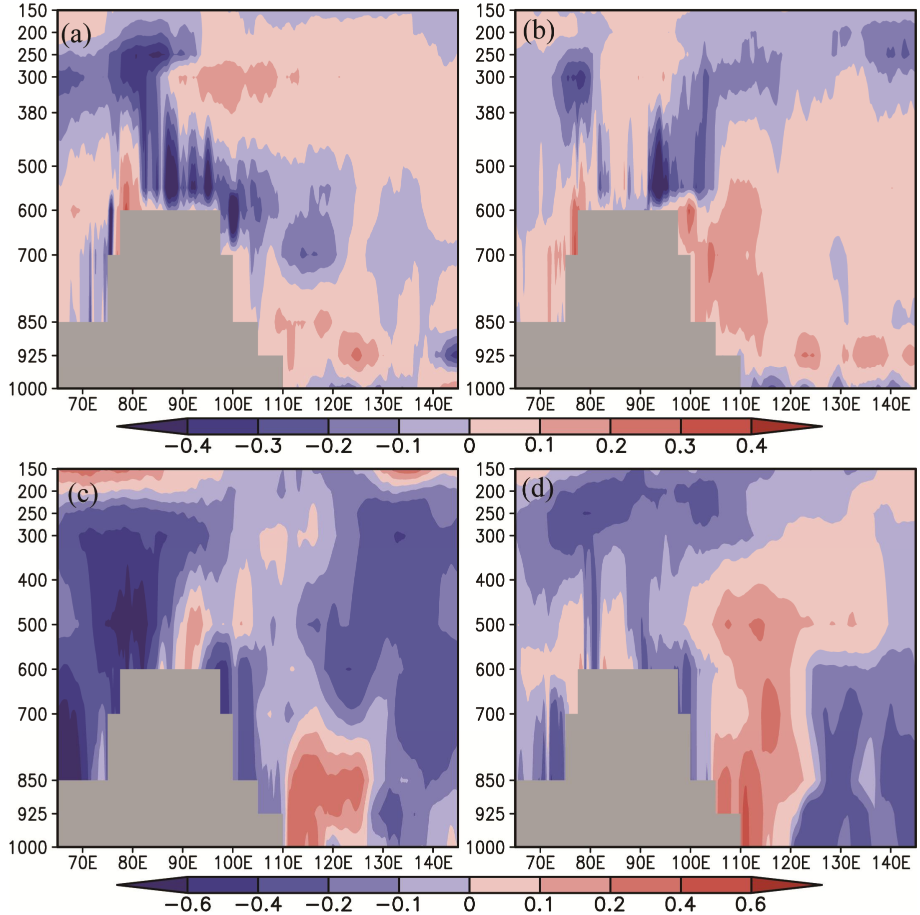

Figure 9: Longitudional cross-section of the differences between CON and SEN averaged over 32–36°N in summer. (a) and (b): net radiative cooling rate (short-wave heating rate + long-wave cooling rate, K day^{-1}) in heavy and light dust years respectively. (c) and (d): as (a) and (b) but for atmospheric temperature.

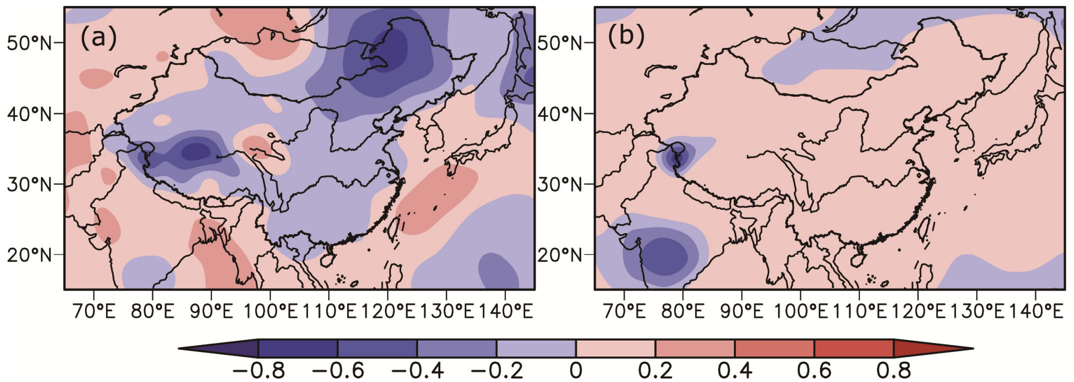

Figure 10: Simulated difference in summer surface temperature between CON and SEN in (a) heavy and (b) light dust years.

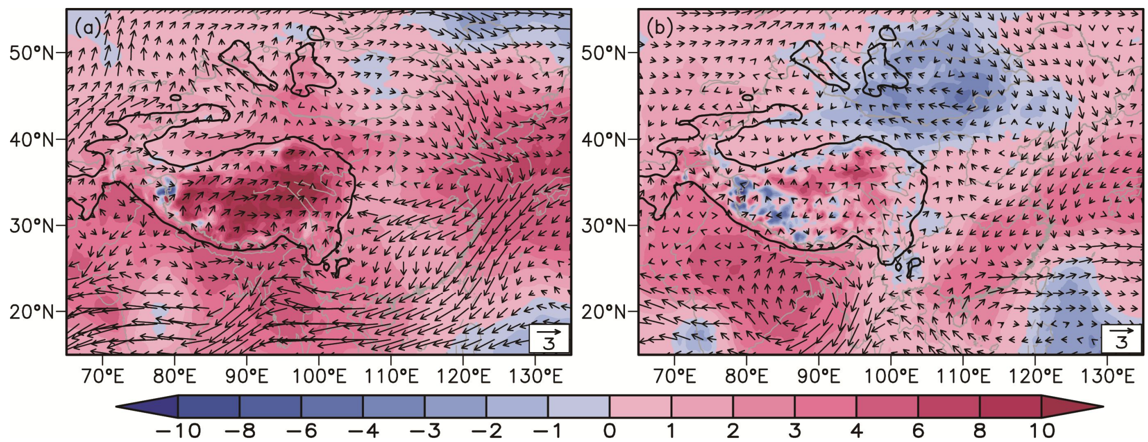

Figure 11: Simulated difference in atmospheric circulation at 850 hPa (vector, m s^{-1}) and geopotential height at 600 hPa (shaded, m) in summer between CON and SEN in (a) heavy and (b) light dust years.

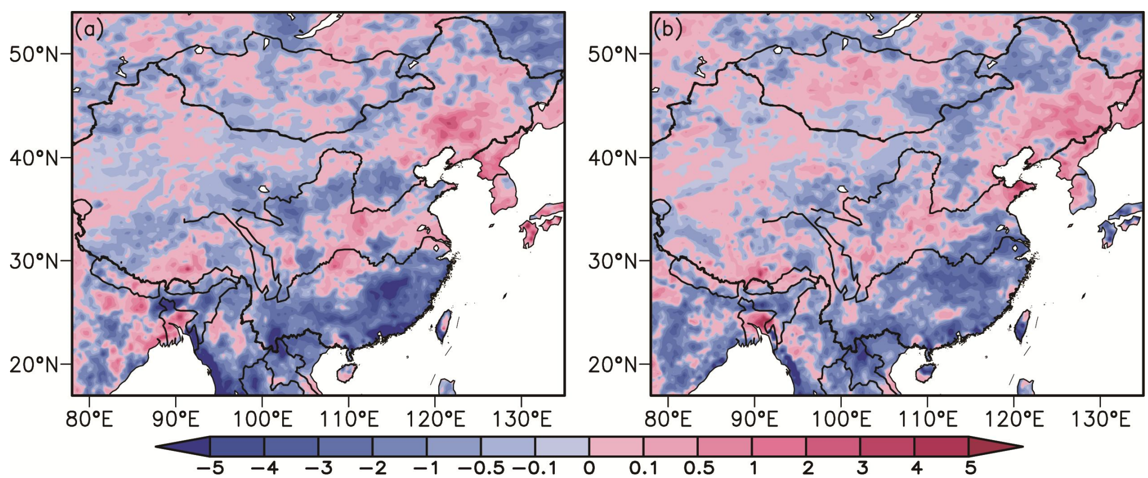

Figure 12: Simulated difference in summer precipitation between CON and SEN in (a) heavy and (b) light dust years.

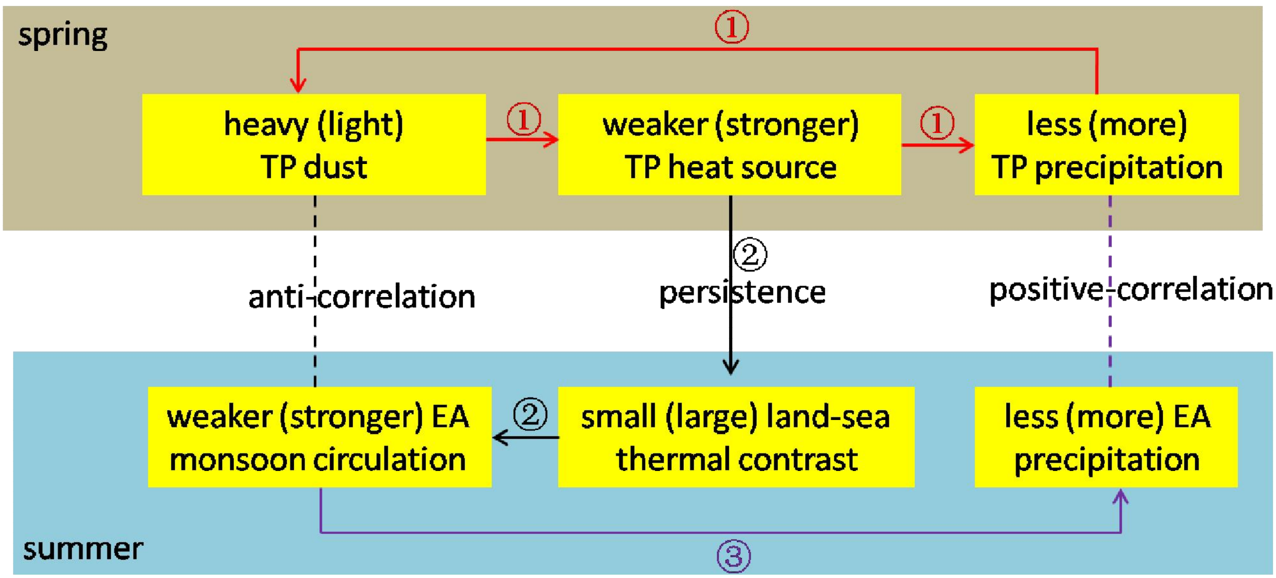

Figure 13: Schematic diagram showing the relation of dust aerosols emitted from the TP in spring with EASM precipitation.