Anonymous Referee #1

Few would have the patience to finish reading an overreaching manuscript that mumbles along.

Many thanks for the patience to read the manuscript and valuable comments.

Global climate change involves diverse aspects from the stratosphere to the ocean, from the polar region to the tropics, and from monsoon to severe storms. Each of these elements should be investigated independently in great detail, but their relationships to each other and their roles in global climate change also warrant investigation. Without the latter, we are unable to see the "big picture". The goal of this study is to provide a framework for putting some of the pieces together by elucidating the connection between the tropical atmosphere and ocean.

We explained this in the introduction of revised paper and also modified the title as "Role of tropical lower stratospheric cooling in deep convective activity".

Major comments 1.
Fig. 6c. Zonal-mean vertical velocity change does not seem robust judging from the vertical structure and the relationship with OLR. (a) Vertical velocity change around 10N, where the authors identified a decrease in zonal mean OLR, changes signs three times within the troposphere. The vertical structure is inconsistent with the first-baroclinic mode structure that dominates the tropical troposphere (e.g., Fig. 6d). (b) Strong upward anomalies between the equator and 5S are inconsistent with overall positive OLR anomalies in the region (Fig. 6a). Something seems wrong here.

Convective activity is largely different according to the geographical situation. In fact, "Strange feature of three times change of signs" in zonal mean vertical velocity around 10° N resulted from a superposition of different structures over the continental and oceanic sectors. The northward shift occurs mainly over the continental sector, but over the oceanic sector, strengthening of vertical velocity occurs around 5° N-10° N without latitudinal shift.

"Strong upward anomalies between the equator and 5° S", is due to a low level convergence over the western Pacific–New Guinea region. This is produced as a pair with the change over the eastern Pacific, a part of the Walker circulation. To facilitate to seeing the association between the change over the western and the eastern Pacific, original Figs. 5 and 10 were merged together and the zonal wind component south of the Equator was added in the revised version.

Then, you may ask why we need the zonal mean. Because, if the forcing is zonal, the lower tropospheric responses largely depend on the underlying geography with zonally asymmetric characteristics. However, above the tropical propose layer (TTL), the zonal mean component of the response becomes significant. Therefore, to investigate the variation related to the stratospheric changes, analysis of the zonal mean field is necessary.

To respond the question, the following text and Fig. A1 (Fig. 8 in revised paper) was added in revised version.

"Therefore, meridional sections of "standardized" mean JAS 1999- 2016 anomalous pressure vertical velocity were calculated for different sectors (Fig. 8). In the case of standardized anomalies of 17-year mean as in Fig. 8, if one assumes a normal distribution, absolute values larger than 0.5 are statistically different from 0 at the 95% confidence level. Top panel shows the zonal mean field which can be comparable to that extracted by the SVD analysis in Fig. 7a. Contours indicate climatology. Middle panels are the same as the top panel, but divided into two parts: (left) African-Asian continental sector (30° W–130°E) and (right) Pacific-Atlantic oceanic sector (130° E–330° E). Strengthening of the upward velocity in the TTL and the lower stratosphere occurs in the continental sector adding to the northward shift in the troposphere, but over the oceanic sector, strengthening of vertical velocity occurs around 5° N-10° N without latitudinal shift. In the oceanic sector, upwelling is largest around 5° N–10° N, and strengthening of the upwelling occurs at the same location as the climatology. If we limit the sector only over the African continent (20° W–20° E) to exclude Indian Ocean influence, the abovementioned continental characteristics becomes further clear (bottom left). Over the oceanic sector, increase of the vertical velocity generally occurs around 5° N (bottom- right), but in the western Pacific sector (130° E– 170° E), upward velocity mainly develops south of the equator (10° S– 0°) (bottom- middle). We also note that in climatological vertical velocity in the western Pacific sector is confined practically in the lower troposphere over the equatorial SH (10° S– 0°). This feature can be attributed to the fact that the convergence of the air occurs over a warm ocean east of New Guinea island (Fig. 3c).

The above results indicate that in spite of the mixture of different profile according sectors, the zonal mean vertical field in the TTL mainly reflects the variation over African-Asian continental sector.

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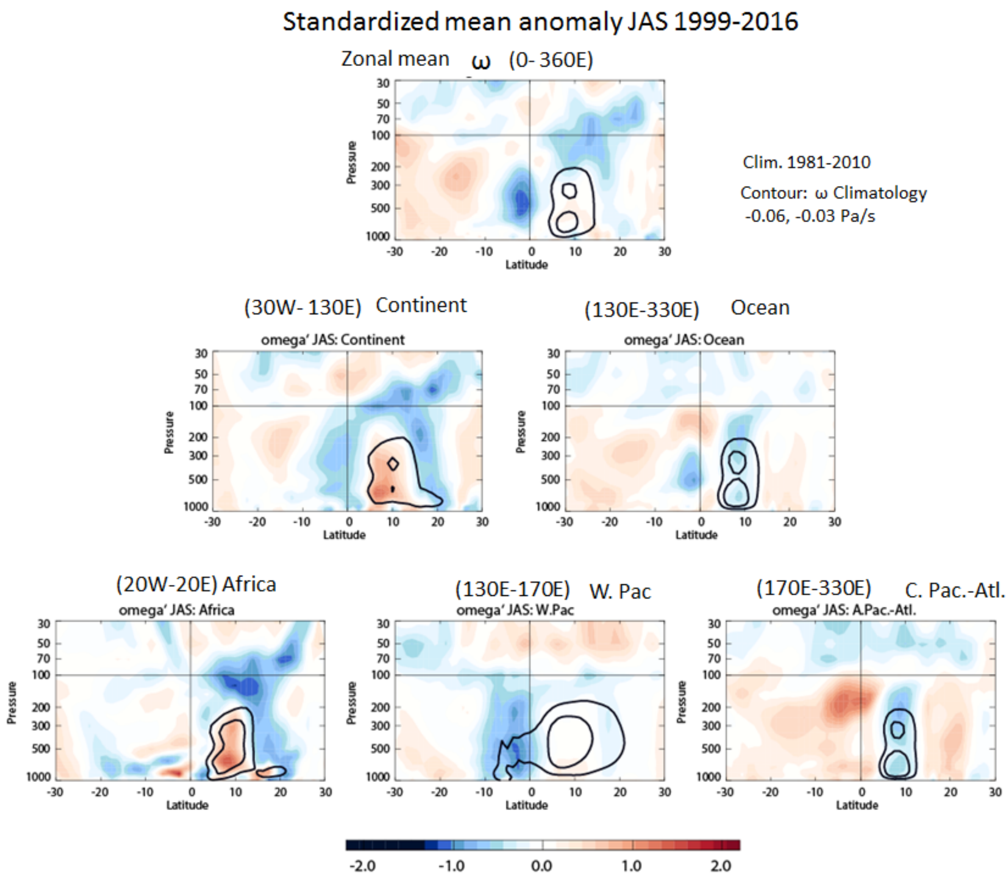

Fig. A1

The same problem applies to the SVD mode showing a strong trend in Fig. 8. The SVD mode with a strong trend features large OLR anomalies in the nonconvective subtropical regions (Figs. 8b,c), raising the question of to what extent the OLR pattern is associated with deep convection and deep vertical motion. Since the focus of the paper is on deep convection, I suggest using precipitation data throughout the paper, instead of OLR.

The reviewer didn't indicate the exact location of "nonconvective subtropical regions". Relatively negative correlation is found in two regions higher than 20° N in original Fig. 8c, one around Pakistan, and the other over Western Pacific (reproduced in the top panel of Fig. A2). It should be noted that strong convection events occur in these region: severe floods in Pakistan in summer 2010 is well known. In the case of western Pacific east of Philippine, deep convection associated with the tropical cyclone frequently occurs (Fig. A2 bottom). Unfortunately, long-term datasets to study the global occurrence frequency of the extreme deep convection does not exist. We will show in Part II of this paper that following a decrease of tropical lower stratospheric temperature, extreme deep convection increased over the western Pacific in association with a development of tropical cyclones.

There are advantages and disadvantages of using precipitation data. Large amount of precipitation is produced by low level convergence, which is of little concern in the present study. We are only interested in precipitation associated with deep convection. Also, direct measurements of global precipitation have limited data record. Thus, precipitation is derived by making use of some assumptions. In this sense, the OLR is better.

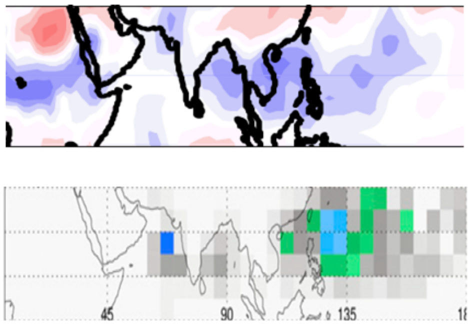

Fig. A2 (top) heterogeneous correction map of OLR, (Bottom) Fraction of deep convective updraft in tropical cyclones (from Fig. 4 of Jiang and Tao, 2014).

Jiang, H. Y., and C. Tao, (2014) Contribution of tropical cyclones to global very deep convection. J. Climate,27, 4313–4336, doi:10.1175/JCLI-D-13-00085.1.

The discussion of the seasonal transition in climatology (Fig. 6, right) is out of blue and the physical relationship to the multi-decadal change in deep convection is ambiguous at the best. I suggest deleting the discussion of the seasonal transition to avoid confusion and streamline the paper.

We agree with the comment, and the discussion on the seasonal transition has been removed.

2. The connection to the Brewer-Dobson circulation is dubious as it currently stands and should be deleted. ACP readers expect robustly tested results, not unsubstantiated speculations.

In this paper we showed a connection between the zonal mean vertical velocity in the tropical lower stratosphere and in the TTL around the rising branch of the Hadley circulation. The stratospheric zonal mean meridional circulation can be designated as Brewer Dobson circulation.

Figure 11 is bizarre: are the narrow bands in vertical velocity in Fig. 11a real, and the meridional dipole in the middle panel of Fig. 11b is averaged out in the meridional mean displayed in the upper panel of Fig. 11a. This is just one example that some speculations do not seem to hold water.

It is difficult to verify whether the narrow bands are real because there are no available observations. However, it is plausible that these features are related to the vertical velocity perturbations near convective overshooting clouds (COV), which penetrate into the TTL beyond the level of neutral buoyancy. Figure A3-top panels show the same vertical velocity field in the original Fig. 11, but the mean anomalous vertical velocity of 1999-2016 is compared with climatological (2007-2017) occurrence frequencies of COV. Increased vertical velocity mainly occurs over the region where COVs are frequent. This relationship is expected because convective overshooting occurs in extreme deep convective clouds, which penetrate into the TTL.

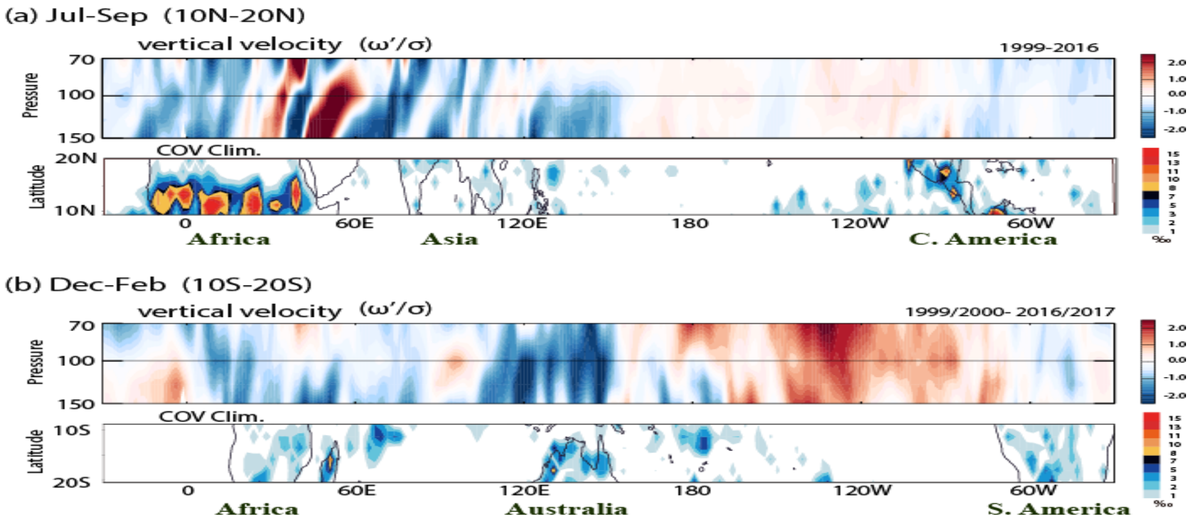

Fig A3

In a previous paper we showed that daily variation of occurrence frequency of the COV during boreal winter is correlated with the zonal mean vertical velocity in the lower stratosphere, while that of the OLR is correlated with the vertical velocity in the troposphere as in Fig. A4.

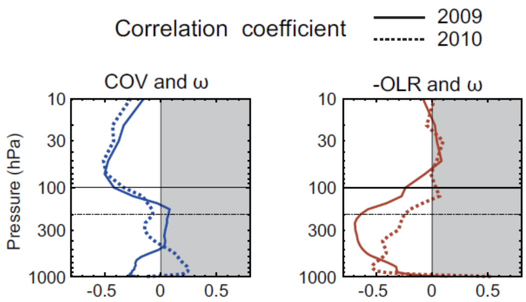

Fig A4 (left) Correlation coefficient between the pressure coordinate vertical velocity at each pressure level and the daily convective overshooting occurrence frequency (COV) averaged over the tropics. (right) Same as the (left) but for the correlation with OLR. [from Kodera et al. (2015), The role of convective overshooting clouds in tropical stratosphere–troposphere dynamical coupling Atmos. Chem. Phys., 15, 6767-6774]

Similar relationship can also be found in climatological field over Africa. Figure A5 compares JAS climatology (2007- 2017 mean) between (left-hand panels) the vertical velocity at 500 hPa and OLR and (right-hand panels) vertical velocity at 100 hPa and the occurrence frequency of the COV over North Africa. At 500 hPa, upwelling extends zonally along 10 °N latitude corresponding to the region of low OLR, whereas upwelling at 100 hPa is broken up in a couple of segments, which roughly corresponds to the region of frequent COV. Similar relationship may also hold on a decadal timescale.

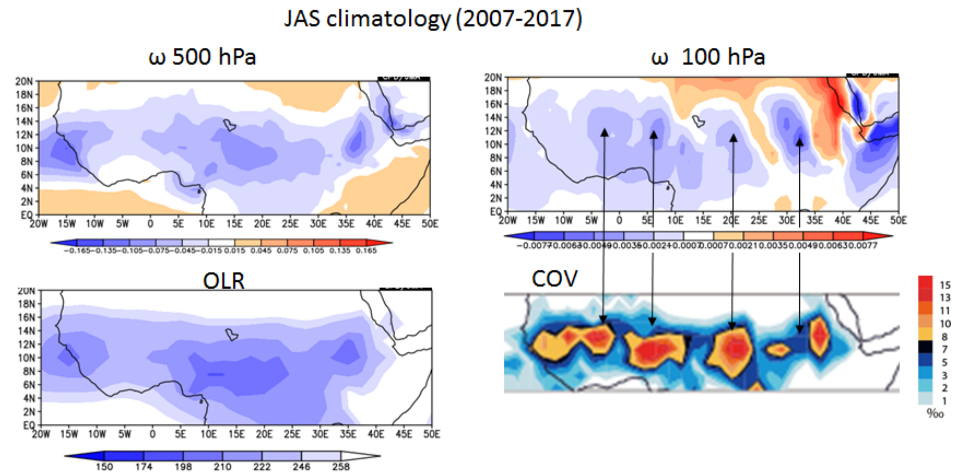

Fig A5

In recent paper, Taylor et al. (2017) reported an increasing trend in deep convective activity over Africa. They showed that the most intense Mesoscale Convective Systems (MCSs) with cloud top temperature lower than −70 °C (air temperature at ~150 hPa) exhibited the largest increasing trend; these MCSs represent clouds penetrating into the TTL. The evolution of these intense MCSs and the pressure vertical velocity at 150 hPa averaged over the similar area is compared in Fig. A6. Rectangles indicate the domains over which averages are calculated for the left panel. There is a quite good correspondence between the evolution of the vertical velocity at 150 hPa and the occurrence frequency of extreme deep convection over Sahel in Africa. It should be noted that the surface precipitation, and lower clouds do not show such clear increasing trend (Taylor et al., 2017). These evidence suggest that the vertical velocities shown in Fig. 11 appear to be related to perturbations induced by the distribution of extreme deep convection.

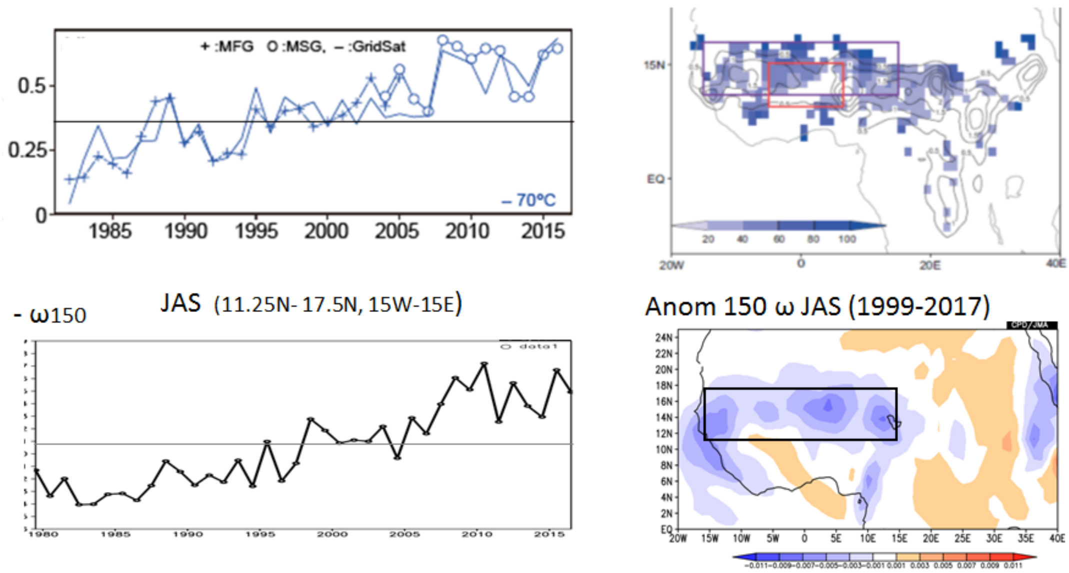

Fig. A6
Top panels: (left) MCS frequency of which cloud top temperature lower than -70 °C. (right) Region of significant trends. The purple rectangle denote the domains used in the left panel. (from Taylor et al., 2017). (Bottom) JAS mean vertical velocity at 150 hPa averaged over the area shown by rectangular in right-hand panel.

Taylor, C. M. et al, 2017: Frequency of extreme Sahelian storms tripled since 1982 in satellite observations. Nature, 544, 475–478.

According to the comment, original Fig. 11 was replaced by Fig. A3 (Fig. 9 in revised paper) and the text was modified as follows in revised version.
" While bottom panels show a distribution of climatological (2007−2017) occurrence frequency of the convective over shooting (COV) in the same latitudinal zone. Inspection reveals that increasing trend of the upwelling occurs over the continental sector, especially where COVs are frequent. This characteristics are commonly seen in both summer hemispheres. The contrast between the continental and oceanic sectors is clearer in the SH where the distribution of lands is simpler. Because the COV occurs in the deep convective clouds penetrating into the TTL beyond the level of neutral buoyancy, such the increased vertical velocity in the TTL over the region of frequent COV seems reasonable. It should also be noted that a connection between the COV and the vertical velocity in the tropical lower stratosphere in day-to-day scale has been shown in a previous study by Kodera et al. (2015)."

3. At the top of page 8, the authors admitted "a discussion of statistical significance : : : is practically impossible". This may be true but the mutual physical relationship among different fields is a minimum test an analysis needs to pass. My major comment 1 questions whether some of the results are robust and physically consistent.

The results of above analyses can be schematically summarized in A7 (Fig. 11) . According to this, we selected 4 key variables which can be considered as fundamental in the recent tropical trends: (a) tropical lower stratospheric temperature in early summer (temperature at 70 h Pa averaged over 20° S–20° N from 16 July-16 August), (b) pressure vertical velocity at the bottom of the TTL (150 hPa) in August, (c) August-October mean "southward" winds south of the equator (10° S–0°) in the western hemisphere (180° W–0°) , (d) tendency in the SST, from the early summer (May-July) to late autumn (October–December) in the tropical Pacific west of South American continent (15° S–5° S, 100° W–80° W). Time series of these 4 variables (a–d) are displayed in A7 (Fig 11)–right.

When all 4 variables become negative (indicated by red dots), we define this as negative event. Similarly when all variables become positive (black dot), it is defined as positive event. All 6 positive events occurred within the first 14 years, while all 7 negative events appeared during the last 13 years. Chi-square test was conducted to examine whether such distribution of the events occurs by chance by dividing the whole 39 years to 3 equal period of 13-year. The result (χ^2 = 23) indicates that such distribution occurs by chance with less than 0.1% of probability. Therefore, there is a statistically significant tendency that negative events occur more frequently in recent decade.

The problem here is, however, whether there is a causal relationship among the variables. We introduced a seasonal variation for the selection of the variable from the stratospheric cooling at the end of July, to a cooling of ocean from summer to autumn. The time evolution tentatively suggests a causality among them, which can be used as working hypothesis, but definite causality needs to be proven in future study. More detailed analysis on selected events will be done to understand causal relationship between the stratospheric and tropospheric change in Part II of this paper.

We added the above text and Fig. A7 (Fig. 11 in revised paper) to revised version.

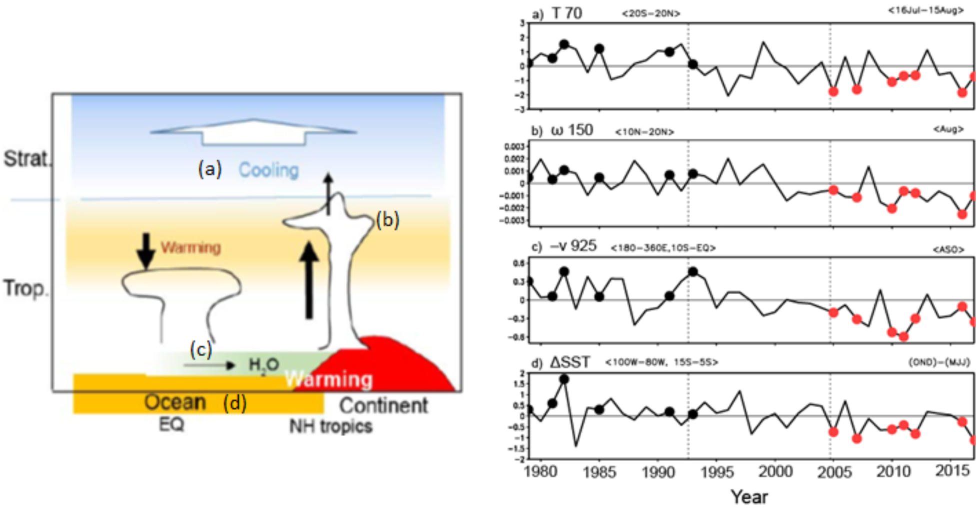

Fig. A7

4. Related to my major comment 1, JAS OLR change (Figs. 6 left, 8-9) is not zonally uniform; specifically the decrease in 10-20N is zonally confined in the African-western Pacific sector. This raises the question of whether the zonal mean even makes sense.

Please see our response to comment 1.

5. If the first three sections are hard to go through, section 4 (Summary and discussion) is impossible to comprehend, full of wild, poorly connected speculations. I urge the authors to summarize robust results that make physical sense first and only then, make some reasonable discussions that would be helpful for future research.

As mentioned earlier in this paper, we have tried to provide a framework for examining the different regions and variables simultaneously to better understand the present global change occurring in the atmosphere-ocean system.

Causal relationship is usually verified by numerical model simulations. However, this requires a model that is capable of realistically reproducing all key processes, none seems existing today. Here, we focused on the validation of one feature of one model - the effects of extreme deep convection on the vertical velocity in the TTL - which is a key element in addressing the present problem.

According to comment, we simplified the discussion section.

Anonymous Referee #2

We would like to thank you for the patience to read the manuscript and give valuable comments. We know it is not usual to investigate climate variation from the ocean surface to the stratosphere in a single paper, but we think it is necessary to get a global view of the recent change.

The authors identified a northward shift of the ITCZ from observations and they related this shift to circulation changes in the stratosphere as a result of stratospheric cooling under global warming. I like the fact that the authors tried a number of correlation analyses between the ITCZ shift and the changes in the tropospheric/stratospheric circulation. However, if I have not missed anything, I found some logical flaws in their arguments and that the mechanism proposed in Figure 13 is not rigorously supported by observations or theories. I wonder if it could be due to a problem of writing and presentation.

Therefore, I recommend a major revision in the first revision round, so that the authors may clarify their ideas, modify the title, write a more in-depth literature review, redo the data analysis, and remove unsupported claims. The following critical comments may be considered during their revision.

Global climate change involves diverse aspects from the stratosphere to the ocean, from the polar region to the tropics, and from monsoon to severe storms. Each of these elements should be investigated independently in great detail, but their relationships to each other and their roles in global climate change also warrant investigation. Without the latter, we are unable to see the "big picture". The goal of this study is to provide a framework for putting some of the pieces together by elucidating the connection between the tropical atmosphere and ocean.

We explain the above in introduction of the paper and also modified the title as Role of tropical lower stratospheric cooling in deep convective activity in revised version.

1. The authors argued that changes in the equatorial stratospheric upwelling drove the ITCZ shift. This argument needs to be supported substantially:

(a) The movement of ITCZ under climate change has been well studied; see Schneider (GRL, 2017, https://doi.org/10.1002/2017GL075817) and references therein. These work have already demonstrated the importance of atmosphere-ocean exchange for determining the location of the ITCZ but the authors have not thoroughly review these observational and theoretical studies. (By the way, the authors mentioned the Hadley cell expansion in Introduction. They should clarify whether the Hadley cell expansion and the ITCZ movement are related.) Whatever the authors have come up with on explaining the ITCZ movements, their explanation (e.g. the stratospheric-led SST changes claimed in this paper) must eventually address those atmosphere-ocean exchange mechanistically and quantitatively.

Because the term ITCZ characterize only low level circulation, we do not use this term in the present study.

The characteristics of the convective zone over the ocean and continent largely differs. Figure A1 shows meridional sections of standardized JAS mean pressure vertical velocity (ω) in different sectors. Top panel is zonal mean ω, while middle panels show average ω by sectors: (left) continental (African-Asian), and (right) ocean (Pacific and Atlantic) sectors. Solid lines indicate climatological vertical velocities (-0.06, -0.03 Pa/s). We can see that the northward shift in the troposphere and large anomalous upwelling around the 100hPa level occur over the land sector. Over the oceanic sector, increase of vertical velocity is limited in the troposphere without changing latitudinal location. Therefore, we consider the study of Schneider on the role of the interaction with the ocean is not so relevant to the present problem of northward shift of the continental convective zone.

Hadley cell expansion mentioned in the paper mainly concerns the descending branch of the Hadley cell as written in introduction, and whether that is related to tropical phenomena like the movement in ITCZ or convection is beyond the scope of this paper. This paper mainly concerns the ascending branch of the Hadley cell around the TTL, which exhibits also a different characteristic from the low level convergence zone.

To indicate the difference of vertical velocity field over the land and ocean sectors, original Fig. 6 was replaced by Fig. A1 (Fig. 8 in revised paper) and the following text were added in the revised version.

"Tropospheric zonal mean vertical velocity shows relatively small connection with the horizontal distribution of the OLR or SST in the SVD analysis in Fig. 7. This can be resulted from the fact that the regional scale variation dominates in the lower troposphere due to the surface geography.

Therefore, meridional sections of "standardized" mean JAS 1999- 2016 anomalous pressure vertical velocity were calculated for different sectors (Fig. 8). In the case of standardized anomalies of 17-year mean as in Fig. 8, if one assumes a normal distribution, absolute values larger than 0.5 are statistically different from 0 at the 95% confidence level. Top panel shows the zonal mean field which can be comparable to that extracted by the SVD analysis in Fig. 7a. Contours indicate climatology. Middle panels are the same as the top panel, but divided into two parts: (left) African-Asian continental sector (30° W–130° E) and (right) Pacific-Atlantic oceanic sector (130° E–330° E). Strengthening of the upward velocity in the TTL and the lower stratosphere occurs in the continental sector adding to the northward shift in the troposphere, but over the oceanic sector, strengthening of vertical velocity occurs around 5° N-10° N without latitudinal shift. In the oceanic sector, upwelling is largest around 5° N–10° N, and strengthening of the upwelling occurs at the same location as the climatology. If we limit the sector only over the African continent (20° W–20° E) to exclude Indian Ocean influence, the abovementioned continental characteristics becomes further clear (bottom left). Over the oceanic sector, increase of the vertical velocity generally occurs around 5 °N (bottom–right), but in the western Pacific sector (130° E–170° E), upward velocity mainly develops south of the equator (10° S–0°) (bottom–middle). We also note that in climatological vertical velocity in the western Pacific sector is confined practically in the lower troposphere over the equatorial SH (10° S–0°). This feature can be attributed to the fact that the convergence of the air occurs over a warm ocean east of New Guinea island (Fig. 3c).

The above results indicate that in spite of the mixture of different profile according sectors, the zonal mean vertical field in the TTL mainly reflects the variation over African–Asian continental sector."

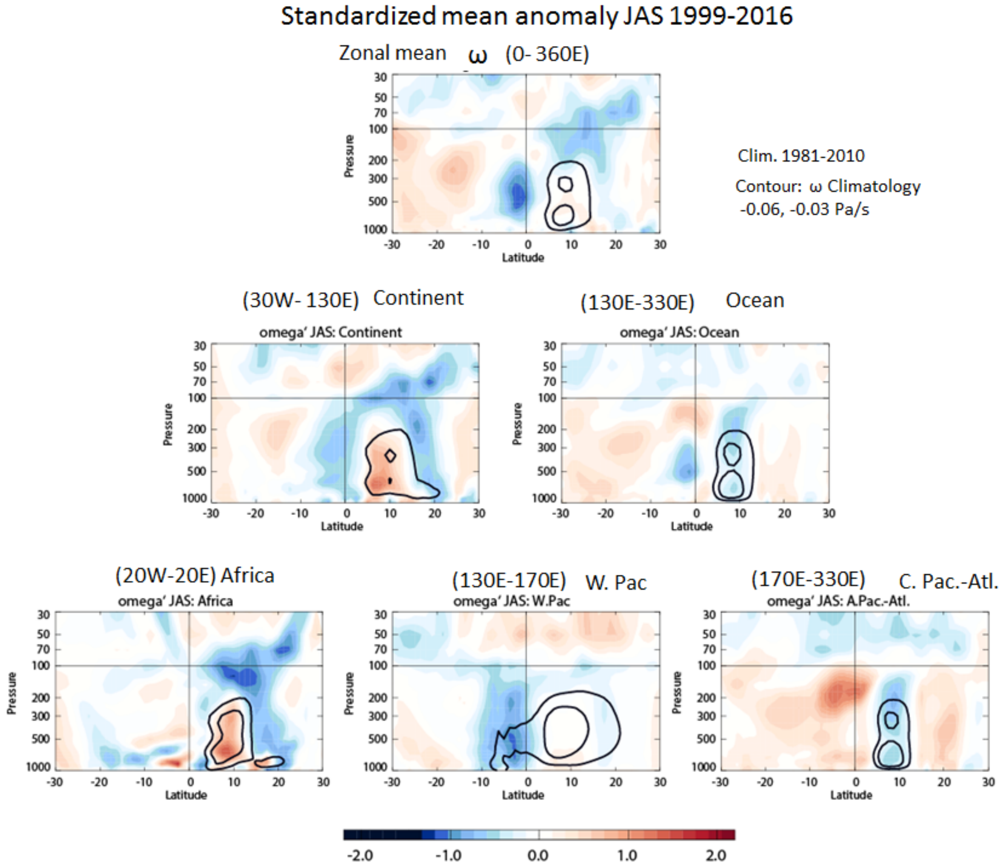

Fig. A1

(b) The logic that they turned their attention to stratosphere is a bit difficult to follow. They first pointed out that changes in convection extended into the tropical tropopause layer. Then they went further to assert that it was the changes in the stratospheric circulation that allowed the convection to get deeper into the tropical tropopause layer. However, there could be a lot of tropospheric causes (e.g. SST changes alone) that made the convection stronger and deeper regardless how the stratosphere changes. Without eliminating all those tropospheric causes first, it is hard to imagine why the stratosphere needs to be involved.

Convective activity depends not only the surface temperature but also that around the cloud top (level of neutral buoyancy). (see Emanuel et al., 2013). Therefore, it is natural to investigate the relationship between the change in deep convective activity and that of the TTL.

Figure A2 a and b are the same SVD analysis as in original Fig. 8b, except for the levels of vertical velocity: (a) 150- 50 hPa around the tropopause, and (b), 1000-300 hPa in the troposphere. It can seen that the zonal mean ω in the troposphere around the rising branch of the Hadley circulation, 15° N-20° N is much closely related to the variation around the TTL than variation in the lower troposphere. According to the comment, original Figure 8 and related text have been replaced by Fig. A2 (Fig. 7 in revised paper) and the following text.

"For this, singular value decomposition (SVD) analysis (Kuroda, 1998) was conducted based on the normalized covariance (correlation) matrix between the zonal mean pressure vertical velocity and the horizontal distribution of the OLR or SST in the tropics (30 °S–30 °N) during boreal summer (JAS) (Fig. 7). Each value at the grid point was weighted by the layer thickness in vertical, and the cosine of the latitude in meridional direction. To investigate the relative importance of zonal mean vertical velocity in different altitude, the SVD calculation are made with the zonal mean pressure vertical velocity of (a) 150–50 hPa and (b) 1000–300 hPa levels. To get a general view of the entire troposphere and stratosphere, the heterogeneous correlation of the vertical velocity is extended to a height range of 1000 to 10 hPa. The levels used for the SVD calculation are indicated by arrows.

In the case of the SVD with tropospheric vertical velocity (Fig. 7b), SVD 1 shows a variation related to the ENSO cycle, while that used the vertical velocity around the TTL (Fig. 7a) ENSO related variation appears as SVD 2. Here we show only those related to the decadal variation. Increasing trends in coefficients are evident in both cases. However, zonal mean ω around 15° N-20° N is much closely related to the variation around the TTL than variation in the lower troposphere. Accordingly, the variation of the OLR around 15° N over African-Asian sector is more closely related with the vertical velocity around the TTL in Fig. 7a. In the case of the SVD with tropospheric vertical velocity (Fig. 7b), heterogeneous correlation with the OLR shows the relationship with convective activity over the equatorial Pacific north of the equator which is confined mainly in the troposphere.

Therefore these results suggest a stronger connection between the convective activity around the rising branch of the Hadley circulation and lower stratospheric circulation through change in the TTL, similar to that suggested in transient response in previous works (Eguchi et al, 2015; and Kodera et al., 2015).

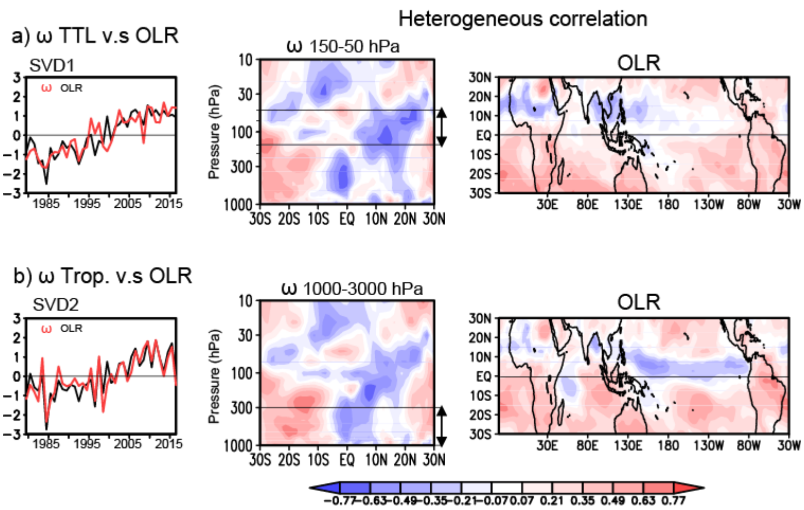

Fig. A2 Time coefficients and heterogeneous correlation maps from SVD analyses. Arrows indicate the level of vertical velocity used for the SDV calculation.

2. All proposed mechanism must be supported quantitatively but there is no detail how much fraction of the ITCZ movement may be caused by the proposed stratospheric changes. In fact, as a scientific paper, there is almost no number in the text that quantify any effects being studied. At most of the time, the authors present some correlation coefficients as hints. But correlations or covariances cannot be used to imply causality between different variables.

Causality should be at least shown by model simulations, which elucidates by how much changes in the stratospheric circulation would cause how much movements of the ITCZ. The models described on page 10 were not designed to isolate the stratospheric effects on the ITCZ movements. In addition, very little has been learned from those models: the model results actually do not quantitatively support their proposed mechanism shown on Figure 13. For this reason, the conclusion of the current manuscript that changes in the equatorial stratospheric upwelling drove the ITCZ shift appears to be speculative. Without a definitive modeling study, the proposed mechanism involving stratosphere, Section 3.4, and Figure 13 are deemed inappropriate.

We acknowledge that quantitative analysis is needed, but many change appear in recent decades in different regions, different seasons, and in different variables. Models can simulate only one aspect, and fail to reproduce changes globally. The goal of this paper is rather to provide a framework for testing these hypotheses in a numerical model, and information how the model should be improved for realistic simulation.

The shift of the convection should be resulted from nonlinear processes. Therefore it may not be possible to linearly separate the portion of the stratospheric influence.

Models are a powerful tool for demonstrating causal relationships, if they are able to reproduce all key processes. In the present case, the model must be able to successfully simulate the effects of extreme deep convection penetrating to the TTL. In this paper, we show that (original Fig. 12) the JMA model has difficulty simulating the vertical velocity in the TTL over land without assimilation of the observational data. This appears to be a common limitation in current models. Until the models are improved, realistic numerical simulation of the decadal variation of extreme deep convective activity discussed in this study may be difficult.

3. Not only the causality should be quantified, the statistical significance of the observed quantities must also be established. Currently, there is no statistical test. Actually, the authors decided to skip the statistical test because "a discussion of the statistical significance of the relationships between variables that exhibit large trends in a short data record is practically impossible". This statement is hard to understand, because (i) if there were large trends in two time series, then it is almost certain that the correlation between these two time series is statistically significant because the noise is relatively small compared to the trends. Did the authors actually mean "weak trends"? And (ii) without establishing the statistical significance, the covariance of the stratospheric and tropospheric variables presented in the manuscript may just be artifacts.

We will make use of the standardized anomalies to demonstrate the statistical significance of the change in vertical velocity: Standardized anomalies of 17-year mean pressure vertical velocity is shown in Fig. A1. In this case, if one assumes a normal distribution, absolute values larger than 0.5 are statistically different from 0 at the 95% significance level. We can also see that the largest difference appears around the tropopause region in the continental sectors.

See also our response to the next question 4.

4. A more serious problem perhaps is their use of the SVD analysis as a way to probe the mechanism. SVD is never intended for causality attribution. SVD can only extract the maximal variability from the data. The 1st mode of the correlation matrix can only be understood as the largest correlation in the tropospheric and stratospheric circulations that share the same secular trend over the past years. (Whether the secular trend is driven by anthropogenic warming is a separate question.) The SVD itself has no implication on whether the stratospheric circulation is driving the tropospheric circulation, or vice versa. They may be independently responding to the same forcing (e.g. anthropogenic warming), and thus the covariance between the

stratospheric and tropospheric trends can be almost certainly statistically significant, as discussed in Comment 3. But such trends do not help to "prove" the proposed mechanism.

Please be noted that we do not try to demonstrate any causality with the SVD: we only use the SVD analysis to extract covariance between two fields and its spatial structures as you mentioned. In revised version, we compare the results of two SVD analysis using different height of the vertical velocity field. This permitted to study which region of vertical velocity is more closely related to the recent trend on the surface.

5. When deriving the proposed mechanism, the authors correlated an index for stratospheric tropical upwelling ($I\omega$) with tropospheric variables such as temperature, SST, and OLR. Figure 9a clearly showed that $I\omega$ was a sum of a decreasing trend plus an oscillatory component. Were the correlation coefficients shown Figure 9b-9d caused by the trend or by the oscillatory component? (The maximal lag of 5 months shown in Figure 9e can only be related to the oscillatory component.) If it is caused by the trend, then, as Comment 4 suggested, the causality is not proven. Indeed, the authors may simply fit each of $\overline{\omega}$, $\overline{T}$, and OLR with a straight line y = c1 * time + c2 and plot c1 to see whether similar Figure 9b-9d could be reproduced. If yes, then there evidence that is the cause of the tropospheric changes.

Generally, year-to-year variation contains variability produced by different sources from that producing decadal trend. For instance, interannual variation in the stratospheric contains stratospheric QBO signal, while tropospheric variation includes ENSO signal. Therefore the correlation study of rapid component by eliminating linear trend may not help for understanding the causal relationship among the trends.

The results of above analyses can be schematically summarized in Fig. 11-left. According to this, we selected 4 key variables which can be considered as fundamental in the recent tropical trends: (a) tropical lower stratospheric temperature in early summer (temperature at 70 h Pa averaged over 20° S–20° N from 16 July–16 August), (b) pressure vertical velocity at the bottom of the TTL (150 hPa) in August, (c) August–October mean "southward" winds south of the equator (10° S–0°) in the western hemisphere (180° W–0°) , (d) tendency in the SST, from the early summer (May–July) to late autumn (October–December) in the tropical Pacific west of South American continent (15° S–5° S, 100° W–80° W). Time series of these 4 variables (a–d) are displayed in Fig 11–right.

When all 4 variables become negative (indicated by red dots), we define this as negative event. Similarly when all variables become positive (black dot), it is defined as positive event. All 6 positive events occurred within the first 14 years, while all 7 negative events appeared during the last 13 years. Chi square test was conducted to examine whether such distribution of the events occurs by chance by dividing the whole 39 years to 3 equal period of 13-year. The result ($\chi^2 = 23$) indicates that such distribution occurs by chance with less than 0.1% of probability. Therefore, there is a statistically significant tendency that negative events occur more frequently in recent decade.

The problem here is, however, whether there is a causal relationship among the variables. We introduced a seasonal variation for the selection of the variable from the stratospheric cooling at the end of July, to a cooling of ocean from summer to autumn. The time evolution tentatively suggests a causality among them, which can be used as working hypothesis, but definite causality needs to be proven in future study. More detailed analysis on selected events will be done to understand causal relationship between the stratospheric and tropospheric change in Part II of this paper.

We added the Fig. A3 and above text to revised version.

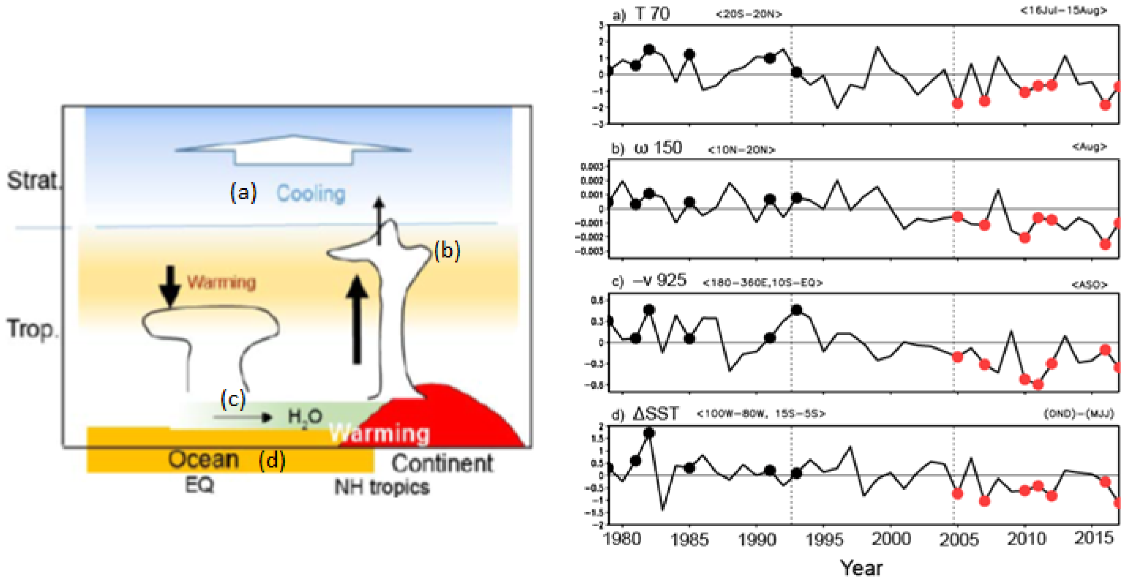

Fig A3.

Role of tropical lower stratospheric cooling in deep convective activity: (I) Recent trends in tropical circulation

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Abstract. Large changes in tropical circulation, in particular those related to the summer monsoon and cooling of the sea surface in the equatorial eastern Pacific, were noted from the mid to late 1990s. The cause of such recent decadal variations in the tropics was studied by making use of a meteorological reanalysis dataset. Cooling of the equatorial southeastern Pacific Ocean occurred in association with enhanced cross-equatorial southerlies, which resulted from a strengthening and poleward shift of the rising branch of the boreal summer Hadley circulation connected to the stratospheric circulation. From boreal summer to winter, the anomalous convective activity centre moves southward following the seasonal march to the equatorial Indian Ocean–Maritime Continent region, which strengthens the surface easterlies over the equatorial central Pacific. Accordingly, ocean surface cooling extends over the equatorial central Pacific. We suggest that the fundamental factor causing the recent decadal change in the tropical troposphere and the ocean is a poleward shift of the rising branch of the summertime Hadley cell, which can result from a strengthening of extreme deep convection penetrating into the tropical tropopause layer, in particular over the continents of Africa and Asia, and adjacent oceans. We conjecture that this effect is produced by a combination of land surface warming due to increased CO_2 and a reduction of static stability in the tropical tropopause layer due to tropical stratospheric cooling. Stratospheric variation has generally been treated as a separated problem from a recent surface climate change, but this framework may be used as working hypothesis for further study.

1 Introduction

Large changes in tropical circulation occurred from the mid to late 1990s. These include a slowdown, or hiatus, of global warming in association with a decrease in the tropical eastern Pacific sea surface temperature (SST) (Kosaka and Xie, 2013; England et al., 2014; Trenberth et al., 2014; Watanabe et al., 2014). Changes were also found in the advancement of the onset of the Asian summer monsoon (Kajikawa et al., 2012; Gautam and Regmi, 2013; Xiang and Wang, 2013; Yun et al., 2014) and an increase in precipitation in West Africa over the Sahel (Fontaine et al., 2011; Brandt et al., 2014; Maidment et al., 2015, Diawara et al., 2016). An increase in precipitation in southern Africa was also noted during the austral summer (Vizy and Cook, 2016). Besides these large-scale circulation changes, changes occurred in mesoscale phenomena such as an increase in Mesoscale Convective Systems (MCSs) over Sahel (Taylor et al., 2017). The tropical cyclone frequency and intensity over the Arabian Sea (Evan and Camargo, 2011) were also reported. Wang et al. (2012) pointed out that these phenomena are, in fact, related to the abovementioned early onset of the Asian summer monsoon. A relationship between tropopause layer cooling and tropical cyclone activity in the Atlantic has also been suggested (Emanuel et al., 2013). Indeed, recent numerical model studies show that cooling of the tropopause impacts the intensity of tropical storms as well as SSTs (Ramsay, 2013; Wang et al., 2014). In this respect, the recent cooling of the tropical tropopause and lower stratosphere from around 2000 (Randel et al., 2006; Randel and Jensen, 2013) should be investigated together with tropical tropospheric change.

The importance of the Pacific Decadal Oscillation (PDO) on decadal changes in global temperature and precipitation has been noted (Meehl et al., 2013; Dong and Dai, 2015; Trenberth, 2015). The most recent hiatus ended around 2013 followed by a large warming due to an El Niño event in 2015 (Hu and Fedorov, 2017; Liu and Zhou, 2017; Urabe et al., 2017; Xie and Kosaka, 2017). However, the El Niño of 2015/16 was different from the previous large 1997/98 El Niño with less warming in the eastern Pacific (Paek et al., 2017), conforming to an increasing trend of the central Pacific-type El Niño (Kao and Yu, 2009; Johnson, 2013). In this sense, the anomalous tropical circulation from the 1990s did not terminate with the hiatus, but is persisting. Similarly, the northward shift of the convective zone in the boreal summer still continues, as shown below.

As a cause of recent tropical changes, multidecadal variations of the atmosphere–ocean coupled mode such as the Atlantic Multidecadal Oscillation have also been proposed (Wang et al., 2013; Kamae et al., 2017). The possible impact of SSTs in different oceanic basins on recent trends in monsoon precipitation was studied by using a coupled ocean model by Kamae et al. (2017). Their results show that the change in Atlantic SST can reproduce recent increasing trends in monsoon rainfall in the northern hemisphere (NH), except for the Asian monsoon. The Atlantic SSTs, however, have practically no effect on the southern hemisphere (SH) such as the African and Australian monsoons. Another difference from observations is that the simulated increase in rainfall occurs mainly over the oceans in low latitudes between the equator and 15° N rather than over continents between around 10° N and 20° N as observed (see Fig. 3 of Kamae et al., 2017). Thus, it is difficult to attribute the recent global trends to a regional mode of decadal oceanic variation alone. In this paper, we first show that the fundamental factor causing the recent decadal trend in the tropics from the mid to late 1990s is not the PDO, but rather a poleward shift of the summertime rising branch of the Hadley circulation connected to the upwelling branch of the stratosphere.

One aspect of the recent trends in tropical circulation that has been reported is the poleward expansion of the tropics (e.g., Davis and Rosenlof, 2012). The expansion of the tropics is rather related to a change in the sinking branch of the Hadley cell driven by midlatitude transient eddies (Kang and Polvani, 2011). There are also other phenomena showing large trends in high latitudes and polar regions that are a part of global change and could be related to tropical change. However, in the interest of brevity, this paper focuses on changes in the tropics.

Global climate change involves diverse aspects from the stratosphere to the ocean, from the polar region to the tropics, and from monsoon to severe storms. Each of these elements should be investigated independently in great detail, but their relationships to each other and their roles in global climate change also warrant investigation. Without the latter, we are unable to see the "big picture". Stratospheric variation has generally been treated as a separated problem from a recent surface climate change. The goal of this study is to provide a framework for putting some of the pieces together by elucidating the connection between the atmosphere and ocean in the tropics.

The paper is organized as follows. After describing the data used for the study in Sect. 2, the results of our analysis are presented in Sect. 3. Finally, a summary and discussion of the causes of the recent tropical changes are presented in Sect. 4.

2 Data

We make use of meteorological reanalysis data produced by the Japan Meteorological Agency (JMA), JRA55 (Kobayashi et al., 2015). A large discontinuity was found at the end of the 1990s in the previous reanalysis product, JRA25, when the TIROS Operational Vertical Sounder (TOVS) on board the National Oceanic and Atmospheric Administration (NOAA) satellite was switched to Advanced TOVS (ATOVS) (Li et al., 2000). This discontinuity has largely been reduced in the JRA55 reanalysis (Kobayashi et al., 2015). To investigate deficiencies in the general circulation model, we make use of a family of JRA55 product: AMIP-type Simulation (JRA55-AMIP), which uses the same model and boundary conditions as JRA55 reanalysis, but without assimilation of observational data in the atmosphere.

Outgoing longwave radiation (OLR) data provided by NOAA are widely used for analysis of convective activity in the tropics. In the present study, we use monthly mean OLR data (1° × 1° latitude–longitude resolution) derived from the High Resolution Infrared Radiation Sounder (HIRS) (Lee et al., 2007), available at ftp://eclipse.ncdc.noaa.gov/cdr/hirs-olr/monthly/. Analysis of the precipitation is performed by making use of Global Precipitation Climatology Project (GPCP) monthly mean data Ver.2.3 (Adler et al., 2003). COBE monthly mean gridded SST data with 1° × 1° boxes compiled by the JMA (Ishii et al., 2005) are used for the study of oceanic surface change.

Climatology is defined by the 30-year mean from 1981 to 2010 in the present study. Seven El Niño events after 1979 are defined by the JMA based on 6-monthly mean SSTs in the Niño 3 sector (5° S–5° N, 150° W–90° W) (available at http://ds.data.jma.go.jp/gmd/tcc/tcc/products/elnino/ensoevents.html). In this study, we define the NH cold seasons 1982/83, 1986/87, 1991/92, 1997/98, 2002/03, 2009/10, and 2015/16 as El Niño winters.

Detection of the tropical overshooting cloud (COV) was made according to the diagnostics developed by Hong et al. (2005), which is based on the brightness temperature differences measured by three high frequency channels of the Advanced Microwave Sensing Unit (AMSU) module B or the Microwave Humidity Sensor (MHS). The data are from NOAA and MetOp satellites: period 2007-2013 for MetOp-A and 2014-2017 for MetOp-B. Their equatorial crossing time is nearly the same (see Fig. 1 of Funatsu et al., 2016). The original data calculated with 0.25°x 0.25° grid was resampled to a coarse one of 2.5°× 2.5° for plotting purposes. Number density of COV is defined as total number of COV detected in each 2.5° x 2.5° bins divided by MetOp-MHS total pixel numbers to remove sampling bias. Units are parts per thousand.

3 Results

3.1 Surface variation vs. OLR

The recent poleward shift of the convergence zone in the boreal summer is identified from the July–September mean anomalous OLR in 1999–2016 from 30-year climatology (1981–2010) (Fig. 1a). A northward shift of the convective activity is seen around the summer monsoon regions over Africa, Asia, and Central America. A poleward shift is further evident in the zonal-mean OLR in Fig. 1b. Climatological OLR peaks around 10° N, while anomalous OLR of the recent period has a maximum around 15° N.

There is a close relationship between the position of the tropical convergence zone and the development of cold tongues in the tropical oceans. Convective activity shifts northward during boreal summer. Accordingly, cross-equatorial winds west of the American and African continents increases which lead to a decrease in the SSTs along the coastal regions during the boreal summer as a part of seasonal cycle. Thus, the primary factor producing cold tongues in the tropical SSTs is the shape of continents, air–sea interaction, and the position of the rising branch of the Hadley circulation, as described in Xie and Philander (1994) and Xie (2004). Therefore, changes in the rising branch of the Hadley circulation, such as can be seen in Fig. 1, can directly impact the equatorial eastern Pacific SSTs.

To investigate whether the northward shift of the convective zone is driven by the PDO, anomalous OLR during the two periods of neutral and negative phases of the PDO is displayed in Figs. 1c and 1d together with anomalous SSTs during those time periods (Figs. 1e and 1f). A characteristic horseshoe pattern in the north Pacific SST is evident during the negative phase of the PDO. Anomalous OLR indicates that the convective activity is enhanced along 15 °N–20 °N latitude irrespective of the phase of the PDO, except for the sector under the direct influence of the PDO in the eastern Pacific, where cooling is larger during the negative phase. However, even during the neutral phase of the PDO, negative anomalies in the SST exist in the tropics west of South America. This suggests that the SST cooling west of South America is not solely driven by the PDO, but is related to the strengthening and northward shift of the convective activity.

The seasonal dependence of the recent trend in tropical convective activity is depicted in Fig. 2. The panels on the left show latitude–time sections of 3-monthly mean anomalous zonal-mean OLR from 1979 to 2016. To better illustrate the evolution in latitudinal structure, the tropical mean (10 °S–25 °N) is further subtracted from the temporal anomaly in Fig. 2. Inspection reveals a change in zonal-mean OLR around 1999 (vertical lines) in all four seasons: The anomalous negative zone shifts northward after 1999 from boreal spring to summer and descends to lower latitudes from boreal summer to autumn following the seasonal march.

The horizontal structure of the anomalous OLR in the recent period (1999–2016) is depicted in the panels on the right for each season: April, May and June (AMJ), July, August and September (JAS), October November and December (OND), and January, February, and March (JFM). The anomalous negative centre of OLR is located around the Maritime Continent during boreal spring, which shifts northward by spreading longitudinally along 15 °N, in particular along the African and Asian sectors during boreal summer. Next, the negative OLR region shifts and clusters over the Indian Ocean region during boreal autumn. At the same time, positive OLR anomalies develop over the equatorial eastern Pacific, suggesting a suppression of convective activity over the Pacific Ocean. During boreal winter, the anomalous convective active region shifts eastward over the Maritime Continent, and suppression of convective activity over the eastern Pacific is sustained during the cold season.

The impact of the recent decadal variation of convective activity on the SST is depicted in Fig. 3. The top panels show the 1999–2016 mean anomalous OLR in (a) JAS, (b) OND. The anomalous horizontal winds at 925 hPa and SSTs are presented in the bottom panels. Because the response of the SST arises following atmospheric circulation, anomalous SSTs during the following month (i.e., August, September and October (ASO) and November, December and January (NDJ) are displayed in Figs. 3c and 3d, respectively. In JAS, the anomalous cross-equatorial flow west of South America intensifies due to a poleward shift of a rising branch of the Hadley cell. The cross-equatorial flow changes from westward to eastward when it crosses the equator according to the change in sign of the Coriolis force. This results in a strengthening of the climatological easterlies in the SH and enhances anomalous convergence near New Guinea. In contrast, easterlies are weakened in the NH, which explains a warming in the north, but a cooling south of the equator. Such a meridional seesaw of the anomalous SSTs and cross-equatorial flow suggests an important role for wind−evaporation−SST (WES) feedback (Xie and Philander, 1994) in recent trends. The centre of anomalous negative OLR moves to the equatorial eastern Indian Ocean from boreal summer to autumn, which results in a strengthening of anomalous easterlies over the equatorial central Pacific and a westward extension of low SSTs over the equator.

In the analysis above, two different features in decadal variability are seen, one over continental and the other over oceanic sectors. Here we examine the characteristics of the variations over the African continent (10° W–40° E) sector and the Pacific Ocean (170° W–120° W) sector, corresponding to the longitudes of Niño 3.4. To identify the different characteristics of convective activity over continents and ocean, the climatological annual cycle in the zonal mean pressure vertical velocity at 300 hPa is depicted in Figs. 4a and 5a. A region of enhanced convective activity migrates north and south over the continent following the seasonal variation of solar heating. It should be noted that the convective zone over Africa shows a jump during the summer monsoon season (Hagos and Cook, 2007). In the case of the Pacific Ocean, the convective zone shows only a small latitudinal excursion and is located in the NH around 5° N throughout the year.

Latitude–time sections of the 3-monthly mean anomalous (departures from climatology) 300 hPa pressure vertical velocity are shown from February 1979 to November 2016 over the African sector in Fig. 4b. The vertical velocity increases from the mid-1990s in both hemispheres around 10°−20° in latitude corresponding to the rising branch of the summertime Hadley circulation in the TTL. Accordingly, the annual mean precipitation over Africa increases in the recent period (1999−2016) in both hemispheres over the Sahel and Namibia.

Over the Pacific Ocean sector (Fig. 5b), strong upward motion appears over the equator when El Niño events occur. The anomalous region of upward motion, however, tends to stay north of the equator after 1999. Accordingly, the annual mean anomalous precipitation during the recent period exhibits a large increase around 5° N−10° N, but decreases over the equator and the SH (Fig. 5c). This change in the atmospheric response to the El Niño event is related to the change in the strength of the cross-equatorial winds (Fig. 5d). After 1999, although SSTs increased over the equator during El Niño, the anomalous northward wind remained stronger and the convective activity tended to stayed in the NH.

Latitude−time section of 3-monthly anomalous SST of the Niño 3.4 sector (Fig. 5f) show little change in latitudinal structure. Figure 5e shows a longitude–time section of the anomalous OLR over the equatorial SH (0°–10° S). The impact of cooling of the eastern equatorial Pacific in the SH can also be seen as a structural change in El Niño/Southern Oscillation (ENSO) phenomena after 1999. Convective activity largely increases over the Pacific during El Niño events before 1999. However, after 1999, Pacific convective activity is suppressed; an increase in convective activity during El Niño is apparent only over the central Pacific. In contrast, convective activity west of 160° E over the Maritime Continent generally increases after 1999. Such changes should be connected with a decadal change in anomalous zonal winds over the tropical SH (10° S−5° N) (Fig. 5f), which can be related to an increased cross-equatorial southerlies through the effect of the Coriolis force.

3.2 S-T coupling

In addition to the change in the troposphere, increasing trend in the tropical lower stratospheric upwelling at 70 hPa, together with the increasing number of sudden stratospheric warming events, was identified by Abalos et al. (2015). A height–time section of the zonal mean temperature around the tropopause, together with the zonally averaged pressure vertical velocity at 70 hPa, shows year-to-year variation of the temperature associated with the quasi-biennial oscillation superimposed on a decadal cooling trend in the lower stratosphere. In the upper troposphere, warming pulses related to El Niño events are superimposed on a warming trend in the troposphere (Fig. 6). A decrease in lower stratospheric temperature and an increase of upper tropospheric temperature lead to a decrease in the static stability of the TTL. It has been suggested that increased stratospheric tropical upwelling tends to enhance deep convective activity, in particular deep penetrating clouds, through change in the static stability of the TTL (Eguchi et al., 2015; Kodera et al., 2015). In the following we will examine how the decadal variation in the stratosphere and troposphere is connected.

For this, singular value decomposition (SVD) analysis (Kuroda, 1998) was conducted based on the normalized covariance matrix between the zonal mean pressure vertical velocity and the horizontal distribution of the OLR or SST in the tropics (30° S–30° N) during boreal summer (JAS) (Fig. 7). Each value at the grid point was weighted by the layer thickness in vertical, and the cosine of the latitude in meridional direction. To investigate the relative importance of zonal mean vertical velocity in different altitude, the SVD calculation are made with the zonal mean pressure vertical velocity of (a) 150–50 hPa and (b) 1000–300 hPa levels. To get a general view of the entire troposphere and stratosphere, the heterogeneous correlation of the vertical velocity is extended to a height range of 1000 to 10 hPa. The levels used for the SVD calculation are indicated by arrows. A variation related to the decadal variation and ENSO cycle appear as first two modes. Here we show only those related to the decadal variation. Increasing trends in coefficients are evident in two cases. However, zonal mean ω around 15° N-20° N is much closely related to the variation around the TTL (Fig. 7a) than variation in the lower troposphere (Fig. 7b). Accordingly, the variation of the OLR around 15° N over African-Asian sector is more closely related with the vertical velocity around the TTL. In the case of the SVD with tropospheric vertical velocity, heterogeneous correlation with OLR show the relationship with convective activity over equatorial Pacific north of the equator which is mainly confined within the troposphere. Therefore these results suggest a stronger connection between convective activity around the rising branch of the Hadley circulation and the lower stratospheric circulation as suggested in previous works (Eguchi et al, 2015; and Kodera et al., 2015).

3.3 Variation over continent and ocean

Tropospheric zonal mean vertical velocity shows relatively small connection with the horizontal distribution of the OLR. This can be resulted from the fact that the regional scale variation dominates in the lower troposphere due to the surface geography. Therefore, meridional sections of "standardized" mean JAS 1999–2016 anomalous pressure vertical velocity were calculated in different sectors (Fig. 8). In the case of standardized anomalies of 17-year mean as in Fig. 8, if one assumes a normal distribution, absolute values larger than 0.5 are statistically different from 0 at the 95% confidence level.

Top panel shows the zonal mean field which can be comparable to that extracted by the SVD analysis in Fig. 7a. Contours indicate climatology. Middle panels are the same as the top panel, but divided into two parts: (left) African-Asian continental sector (30° W–130°E) and (right) Pacific-Atlantic oceanic sector (130° E–330° E). Strengthening of the upward velocity in the TTL and the lower stratosphere occurs in the continental sector adding to the northward shift in the troposphere, but over the oceanic sector, strengthening of vertical velocity occurs around 5° N-10 °N without latitudinal shift. In the oceanic sector, upwelling is largest around 5° N–10° N, and strengthening of the upwelling occurs at the same location as the climatology. If we limit the sector only over the African continent (20° W–20° E) to exclude Indian Ocean influence, the abovementioned continental characteristics becomes further clear (bottom left). Over the oceanic sector, increase of the vertical velocity occurs around 5 °N (bottom- right), but in the western Pacific sector (130° E– 170° E), upward velocity mainly develops south of the equator (10° S– 0°) (bottom- middle). We also note that in climatological vertical velocity in the western Pacific sector is confined practically in the lower troposphere over the equatorial SH (10° S– 0°). This feature can be attributed to the fact that the convergence of the air occurs over a warm ocean east of New Guinea island (Fig. 3c). This result indicates that in spite of the mixture of different profile according sectors, the zonal mean vertical field in the TTL mainly reflects the variation over African-Asian continental sector.

Continuity in a zonally averaged field does not necessarily mean actual continuity at each location as shown in the above analysis. To investigate more in detail the continuity within the rising branch of the Hadley circulation from the upper troposphere to the stratosphere, longitude−height sections of the normalized anomalous pressure vertical velocities averaged over 10°−20° latitudes in the summer hemisphere are displayed in top panels in Fig. 9 a and 9b. While bottom panels show a distribution of climatological (2007−2017) occurrence frequency of the convective over shooting (COV) in the same latitudinal zone. Inspection reveals that increasing trend of the upwelling occurs over the continental sector, especially where COVs are frequent. This characteristics are commonly seen in both summer hemispheres. The contrast between the continental and oceanic sectors is clearer in the SH where the distribution of lands is simpler. Because the COV occurs in the deep convective clouds penetrating into the TTL beyond the level of neutral buoyancy, such the increased vertical velocity in the TTL over the region of frequent COV seems reasonable. It should also be noted that a connection between the COV and the vertical velocity in the tropical lower stratosphere in day-to-day scale has been shown in a previous study by Kodera et al. (2015).

To get insight into stratosphere-related variation in the troposphere, the JAS mean vertical pressure velocity (ω) at 30 hPa averaged over tropical SH (0°–25° S) is chosen as the index of stratospheric mean meridional circulation (I_ω) (Fig. 10a). The correlation coefficient between I_ω and zonal mean ω at each grid point (Fig. 10b) shows a correlation pattern very similar to the SVD mode in Fig. 7a. To indicate the relationship between the climatological residual mean-meridional circulation, the mass-weighted stream function diagnosed by the method of Iwasaki (1992) is displayed by contours in Figs. 10b and 10c. We can see that the variation in stratospheric upwelling in the Brewer–Dobson circulation is connected to the Hadley circulation in the TTL around 15° N.

The correlation between I_ω with zonal-mean temperature at each grid point from the 90° S to 90° N, is shown in Fig. 9c. The tropical upwelling is not only related to a cooling in the tropics and the summer hemisphere, but it is also related to a warming in the downwelling region around the winter polar stratosphere (Fig. 10c). This illustrates the dynamical nature of the recent tropical stratospheric cooling. Stratospheric upwelling is also connected with convective activity along 15° N−20° N (Fig. 10d), as already discussed above. The correlation coefficients between I_ω and 925 hPa zonal and meridional winds at each grid point are shown by arrows in Fig. 10e. An increase in the cross-equatorial winds in the eastern Pacific and Atlantic is observed. The impact of near surface wind variation on the SST can be seen in the lagged correlation with the SST in Fig. 10e. The cooling in the equatorial eastern Pacific is largest with a time lag of 5 months (i.e., December, January and February (DJF), consistent with the development of La Niña–like SSTs from boreal autumn in Fig. 3.

The results of above analyses can be schematically summarized in Fig. 11–left. According to this, we selected 4 key variables which can be considered as fundamental in the recent tropical trends: (a) tropical lower stratospheric temperature in early summer (temperature at 70 h Pa averaged over 20° S–20° N from 16 July-16 August), (b) pressure vertical velocity at the bottom of the TTL (150 hPa) in August, (c) August-October mean "southward" winds south of the equator (10° S–0°) in the western hemisphere (180° W–0°) , (d) tendency in the SST, from the early summer (May-July) to late autumn (October–December) in the tropical Pacific west of South American continent (15° S–5° S, 100° W–80° W). Time series of these 4 variables (a–d) are displayed in Fig 11–right. When all 4 variables become negative (indicated by red dots), we define this as negative event. Similarly when all variables become positive (black dot), it is defined as positive event. All 6 positive events occurred within the first 14 years, while all 7 negative events appeared during the last 13 years. Chi-square test was conducted to examine whether such distribution of the events occurs by chance by dividing the whole 39 years to 3 equal period of 13-year. The result (χ^2 = 23) indicates that such distribution occurs by chance with less than 0.1% of probability. Therefore, there is a statistically significant tendency that negative events occur more frequently in recent decade.

The problem here is, however, whether there is a causal relationship among the variables. We introduced a seasonal variation for the selection of the variable from the stratospheric cooling at the end of July, to a cooling of ocean from summer to autumn. This time evolution tentatively suggests a causality among them, however, more exact study is needed to really demonstrate.

4. Summary and discussion

Characteristics of the recent tropical circulation change from the middle to the end of the 1990s can be summarized as follows. Decadal cooling in the eastern Pacific occurs in relationship with an increase in cross-equatorial winds, and the easterlies in the tropical SH, which is also related with a strengthening of the convective activity over the African-Asian sector during the boreal summer (Fig. 3). Also, a correlation analysis (Fig. 10) indicates that these variation in the convective activity and the SSTs are related with the vertical velocity around the tropopause levels.

Although it is difficult to demonstrate the causal relationship statistically among the variables exhibiting large trends, such as (a) lower stratospheric temperature, (b) upwelling in the TTL, (c) cross-equatorial winds near surface, and (d) SST tendency from boreal summer to autumn in Fig. 11, they could be related by following processes: The relationship between the convective activity in the rising branch of the Hadley circulation and the cooling of the tropical eastern Pacific can be explained according to the work of Xie (2004) involving a wind−evaporation−SST (WES) feedback. Connection between the tropical stratospheric cooling and the extreme deep convection around rising branch of the Hadley cell in summer hemisphere has been suggested in previous studies (Eguchi et al., 2015; Kodera et al., 2015). Therefore, a combination of these two processes can be used as a "working hypothesis" for the recent tropical change as in Fig. 11.

Although the period of observation may be too short, Aumann and Ruzmaikin (2013) reported that the tropical deep convection over land shows an increasing trend, while that over oceans shows an decreasing trend based on 10 years of Atmospheric Infrared Sounder (AIRS). Furthermore, Taylor et al. (2017) showed that intense mesoscale convective system of which cloud top temperature lower than -70° C largely increased over the Sahel since 1982. The temperature of −70° C corresponds approximately to the air temperature at 150 hPa. This means that the extreme deep convection penetrating to the TTL largely increased over African continent consistent with the present study.

For the study of the causal relationship, numerical model simulation is usually performed. For this, however, model should be able to reproduce correctly involved key processes, which should be extreme deep convection penetrating to the TTL in the present case. To examine the model performance, we compare the climatological mean vertical velocities around the TTL from the JRA55 reanalysis and the JRA55-AMIP run. The latter uses the same atmospheric general circulation model and boundary conditions (SST and sea ice) as JRA55, but without the assimilation of observed data (Kobayashi and Iwasaki, 2016). Usually, the difference arises from the model deficiency. In JRA55-AMIP, upwelling in the summer tropics and downwelling in the winter polar stratosphere are weaker (Fig. 12a). The horizontal distribution of the difference in pressure vertical velocity at 150 hPa (lower panels in Fig. 12a) show that model upwelling is weaker in summer over land and adjacent oceans around 10°–20° in latitude. Regions of weak upwelling in the model coincide with the regions of increasing recent decadal trends in upwelling. Because the observed change is rather opposite to the model biases, it is possible that current models may not properly simulate the recent trends in the tropics.

Concerning to the origin of trend, the increasing trend in the Earth's surface temperature is generally attributed to the increase in greenhouse gases like CO_2 (IPCC, 2013). Such a change in the radiative forcing may explain the global characteristics of the recent change. The effect of increased CO_2 can be divided into a direct radiative effect and an indirect effect through changes in the SST. Model experiments have shown that the direct radiative effect of CO_2 increases tropical upward motion particularly over the Sahelian sector, whereas it suppresses upwelling over the oceanic sector in the Pacific (Fig. 8 of Gaetani et al., 2016). An increase in CO_2 raises the Earth's surface temperature, but also decreases stratospheric temperatures. Note, however, that the recent cooling in the lower stratosphere–tropopause region is also due to a dynamical effect (Abalos et al., 2015). The sudden decrease in lower stratospheric water vapour around the year 2000 (Randel et al., 2006; Hasebe and Noguchi, 2016) suggests a dynamical forcing of the change. Further investigation is needed to determine whether the stratosphere is merely passively responding to the tropospheric warming or playing an active role in the tropospheric circulation change.

Studies of long-term trends face problems that often arise from changes in the observation systems, stations or the drift of the orbit of the satellites. Therefore, to avoid these problems and to better understand the stratosphere–troposphere coupling processes, we conducted additional case studies on the northward shift of convective activity similar to that associated with the recent trend, but associated with a sudden cooling of the tropical stratosphere during a boreal summer by using convective overshooting and cloud top data derived from recent satellite observations (paper in preparation).

5 Data availability

Datasets used in this paper are all publicly available. Meteorological reanalysis datasets created by JMA, JRA55, and JRA-AMIP are available from http://search.diasjp.net/en/dataset/JRA55 and http://search.diasjp.net/en/dataset/JRA55_AMIP, respectively. The COBE monthly mean SST dataset can be obtained from the JMA website (http://ds.data.jma.go.jp/tcc/tcc/products/elnino/cobesst/cobe-sst.html). Monthly mean HIRS OLR data can be obtained from NOAA by FTP (ftp://eclipse.ncdc.noaa.gov/cdr/hirs-olr/monthly/). The GPCP monthly mean precipitation dataset is obtained from the NOAA website (https://www.esrl.noaa.gov/psd/data/gridded/data.gpcp). AMSU/MHS data are available at NOAA's Comprehensive Large Array Data Stewardship System. In this work, AMSU/MHS raw data were obtained with support from the INSU-CNES French Mixed Service Unit ICARE/climserv/AERIS and accessed with the help of ESPRI/IPSL.

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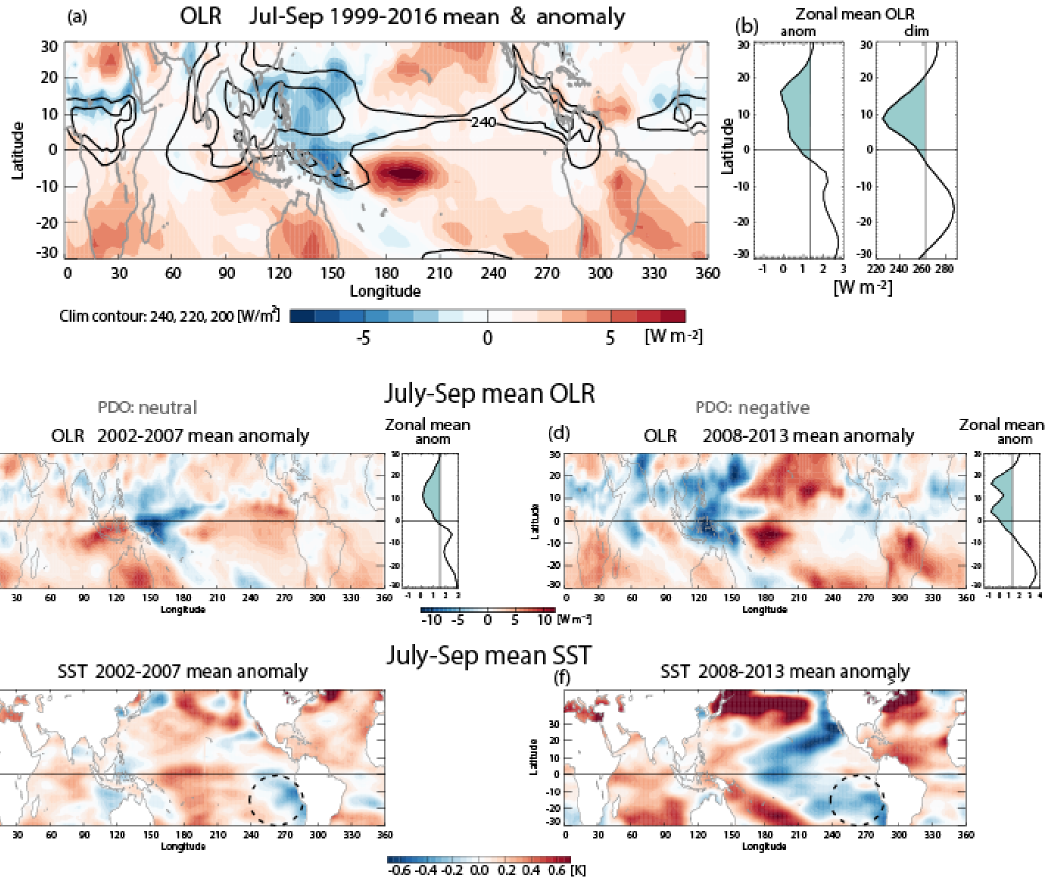

Figure 1: (a) Climatological (1981-2010) July–September mean OLR (contours: 240, 220, and 200 W m^{-2}) and anomalous July–September OLR (departures from climatology) in 1999-2016 (color shading). (b) Zonal mean profiles of (a): anomalies from climatology (left) and climatology (right). (c) Anomalous OLR as in (a) and (b, left) but for 2002-2007 period. (d) Anomalous OLR as in (a) and (b, left) but for 2008-2013 period. (e,f) Anomalous July–September SST (departures from climatology) in 2002-2007 (e) and 2008-2013 (f).

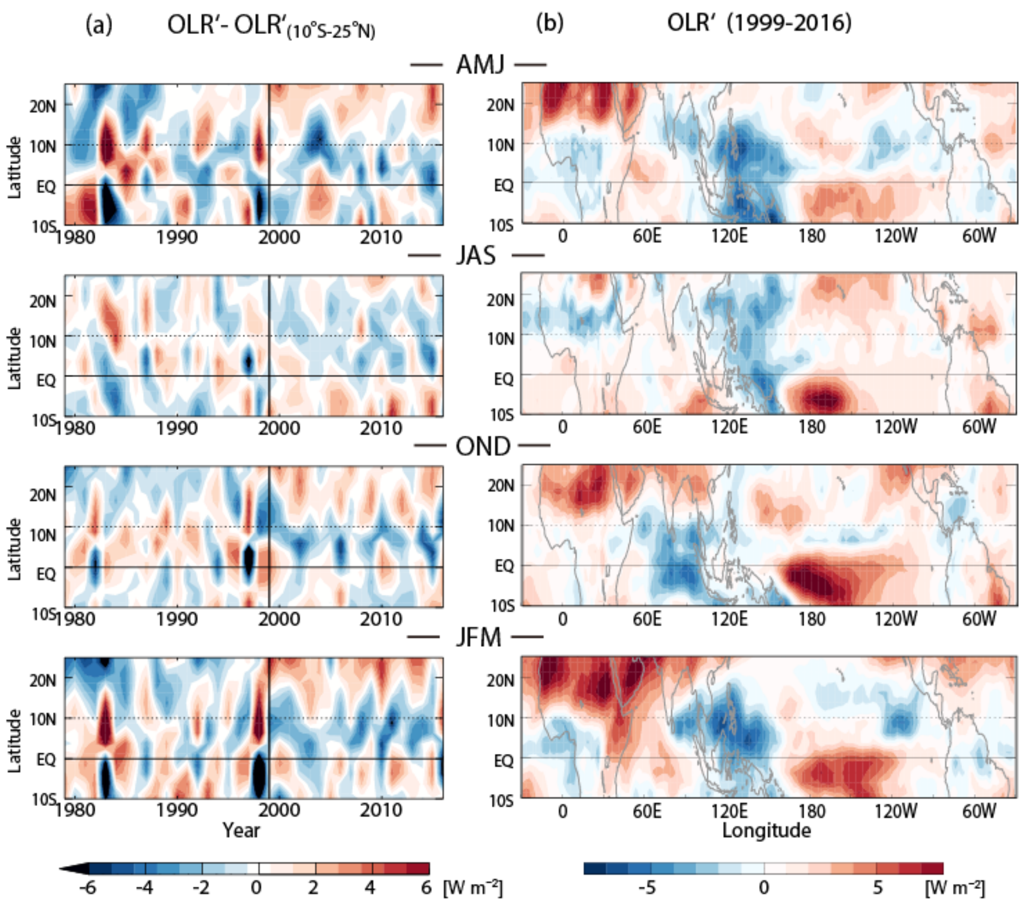

Figure 2: (a) Latitude–time sections of 3-month seasonal mean, zonally-averaged anomalous (departures from 1981-2010 climatological tropical 10° S–25° N mean) OLR in 1979-2016. Vertical solid lines indicate the year 1999. (b) Horizontal distributions of 3-month seasonal mean anomalous OLR from climatology during 1999/2000–2015/2016 period.

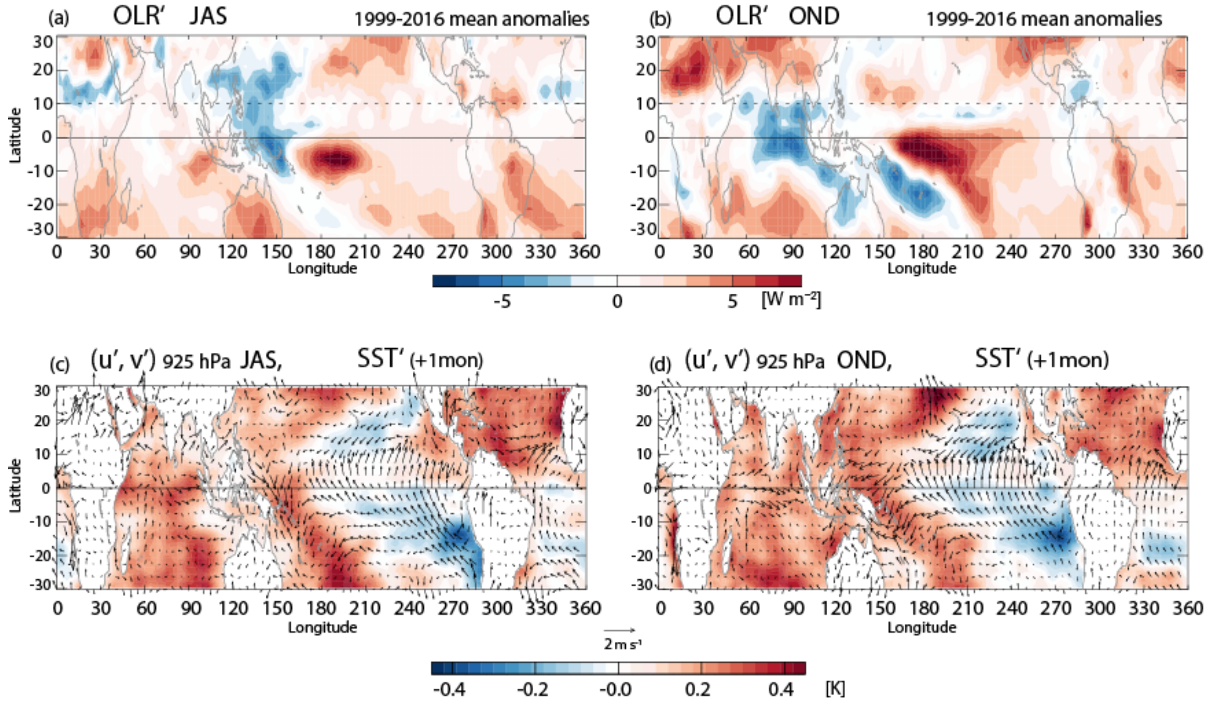

Figure 3: (a) JAS and (b) OND mean anomalous OLR for the period 1999-2016. (c) JAS and (d) OND mean anomalous horizontal winds at 925 hPa (arrows) for the period 1999–2016 superimposed on anomalous SSTs (color shading) with a one-month lag (i.e., ASO and NDJ, respectively).

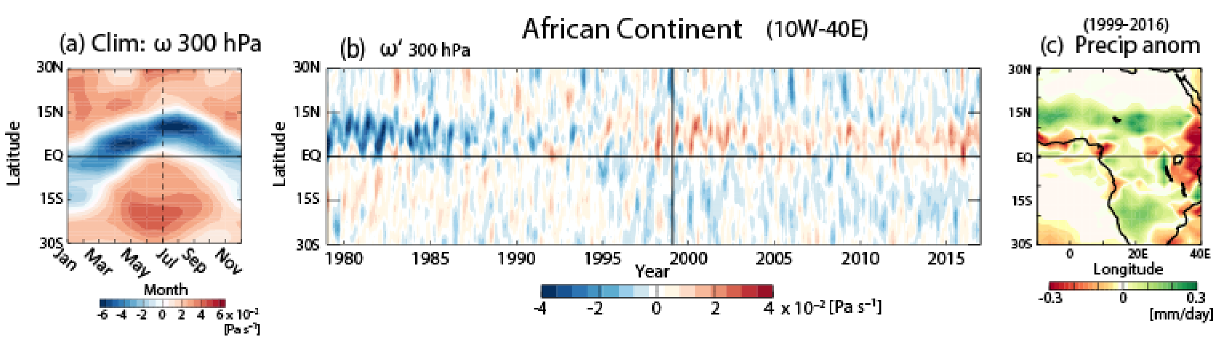

Figure 4: (a) Latitude–time section of the climatological zonal-mean pressure vertical velocity at 300 hPa averaged over the African sector (10° W–40° E). (b) Latitude–time section of monthly mean anomalous pressure vertical velocity from February 1979 to November 2016. (c) Latitude–longitude map of annual mean anomalous precipitation during 1999–2016 over the African sector. Three monthly running mean has been applied on data prior to the presentation for (b).

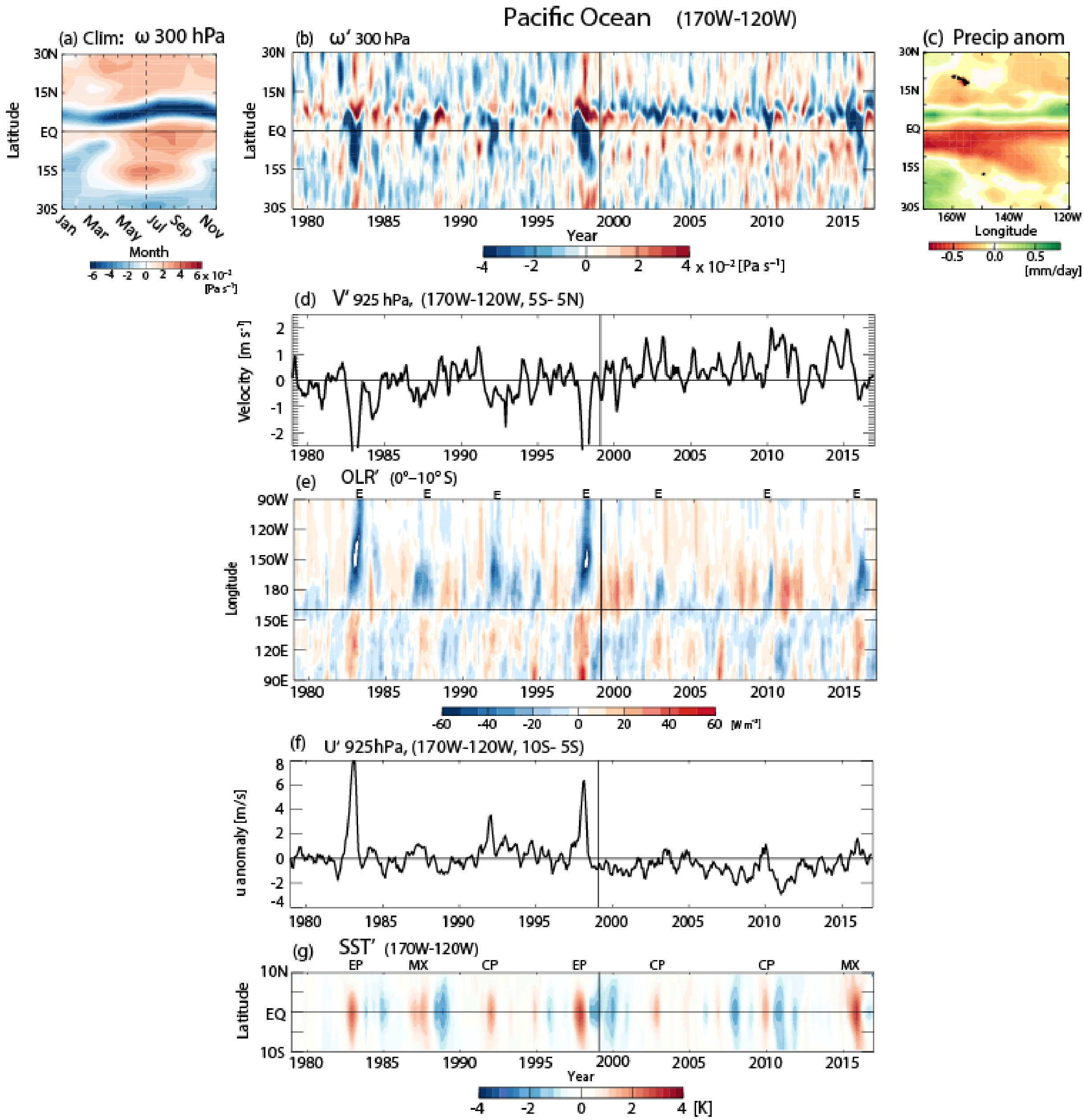

Figure 5: (a- c) Same as in Fig. 4 except for the eastern Pacific, Niño 3.4 (170° W–120° W) sector. (d) Monthly mean anomalous meridional wind component around the Equator (5° S–5° N) over the Niño 3.4 sector. (e) Similar to (a), except for the time-longitude section of OLR around the Equator (5° S–5° N) over the Indian Ocean–Pacific sector. (f) Same as (d), except for zonal wind component around tropical SH (10° S–5° S). (g) Monthly mean anomalous SST over the Nino 3.4 sector. "EP", "CP", and "MX" indicate different types of El Niño events (Peak et al., 2017): EP, eastern Pacific; CP, central Pacific; MX, mixed type.

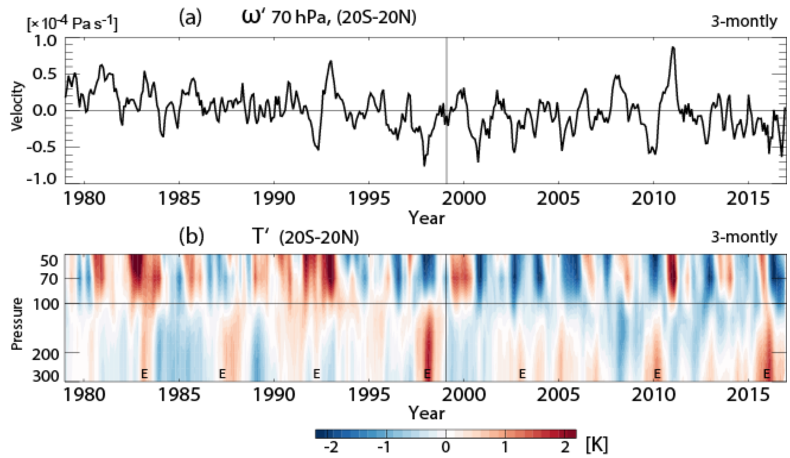

Figure 6: (a) 3-month mean anomalous pressure vertical velocity in the tropics (20° S–20° N) at 70 hPa. (b) Height–time section of 3-month mean anomalous temperature in the tropics (20° S–20° N). "E" denotes an El Niño event.

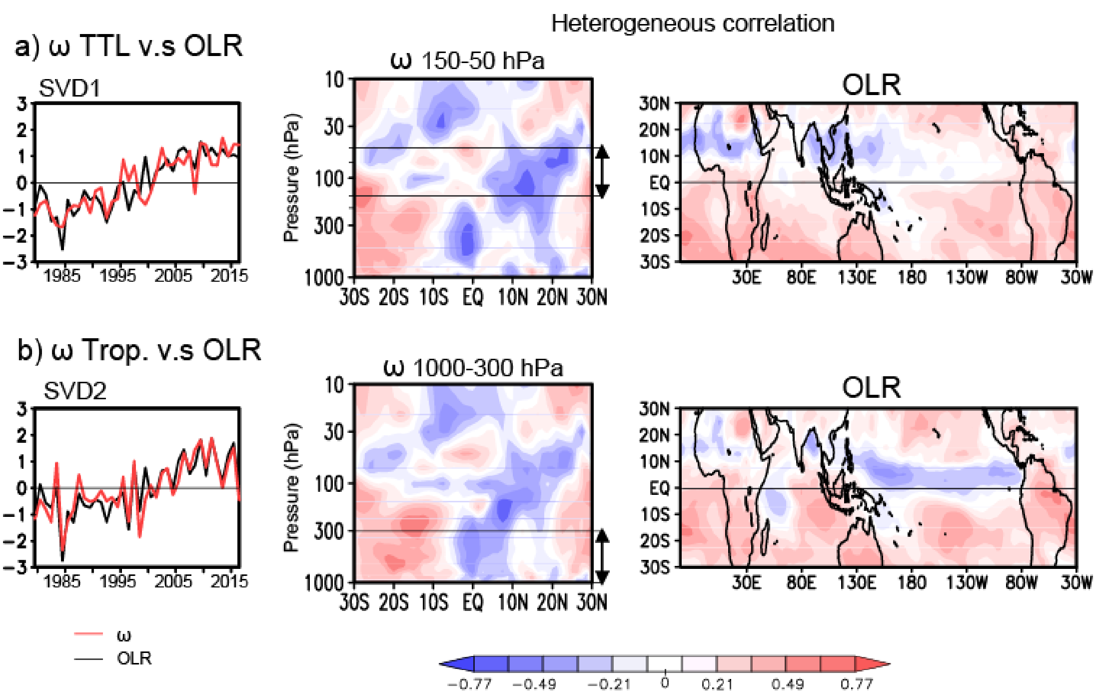

Figure 7: SVD analysis of the zonal mean anomalous pressure velocity and anomalous OLR (0°–360° E) in the tropics (30° S–30° N) during JAS from 1979 to 2016. (a) SVD1 of pressure velocity around the TTL (50–150 hPa) (b) SVD2 of pressure velocity in the troposphere (1000–300 hPa). (from left to right) time coefficients, heterogeneous correlation of OLRs, and heterogeneous correlation map of the pressure velocity extended to 1000–10 hPa. Arrows indicate the levels used for the calculation of the SVD.

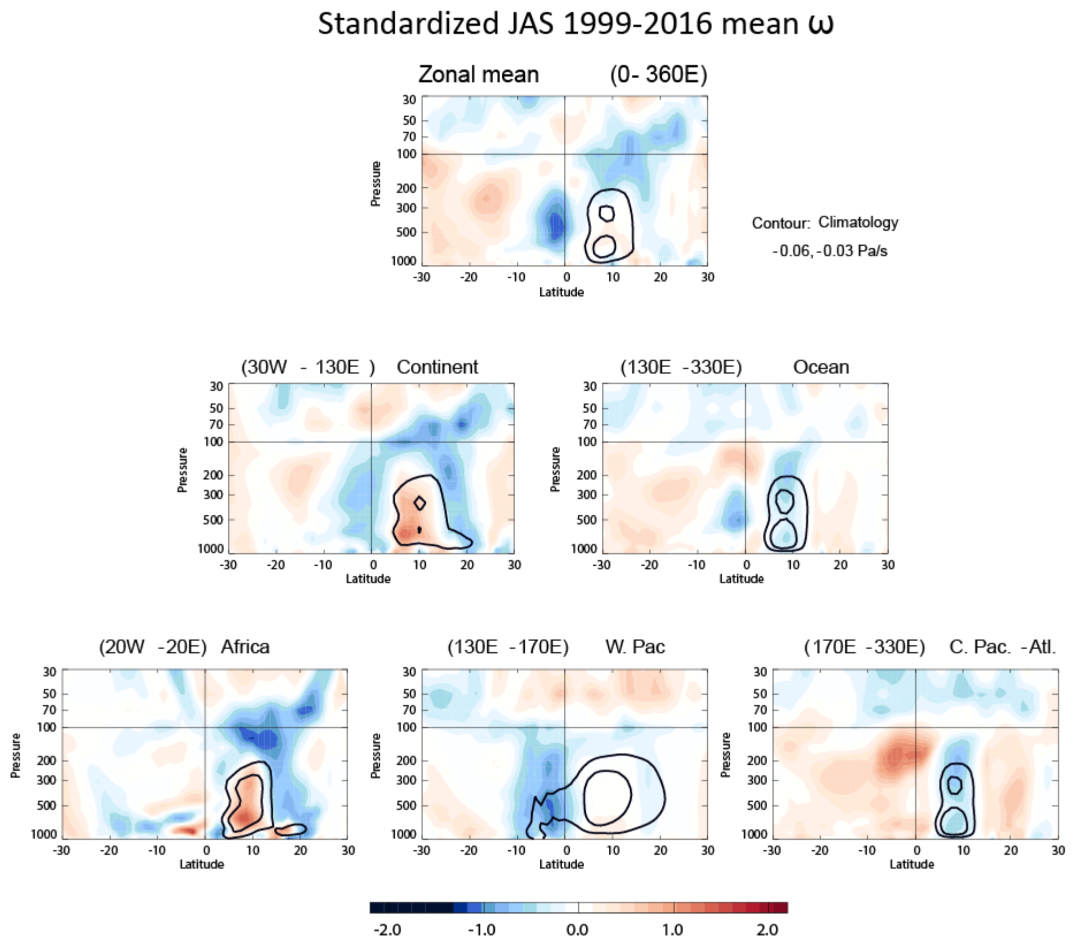

Figure 8: (Top) Standardized anomalous OLR in JAS 1999-2016 (departures from 1981-2010 climatology). Climatological JAS mean, zonal mean pressure velocity is indicated as contours (–0.06 and–0.03 Pa s^{-1}). Middle panels: Same as the top panel, except for (left) African-Asian continental sector (30° W–130° E) and (right) Pacific-Atlantic oceanic sector (130° E–330° E). Bottom panels: Same as the top panels, but for (left) African continent (20° W– 20° E) sector, (middle) Western Pacific sector (130° E–170° E), and (right) Central Pacific–Atlantic (170° E– 330° E) sector.

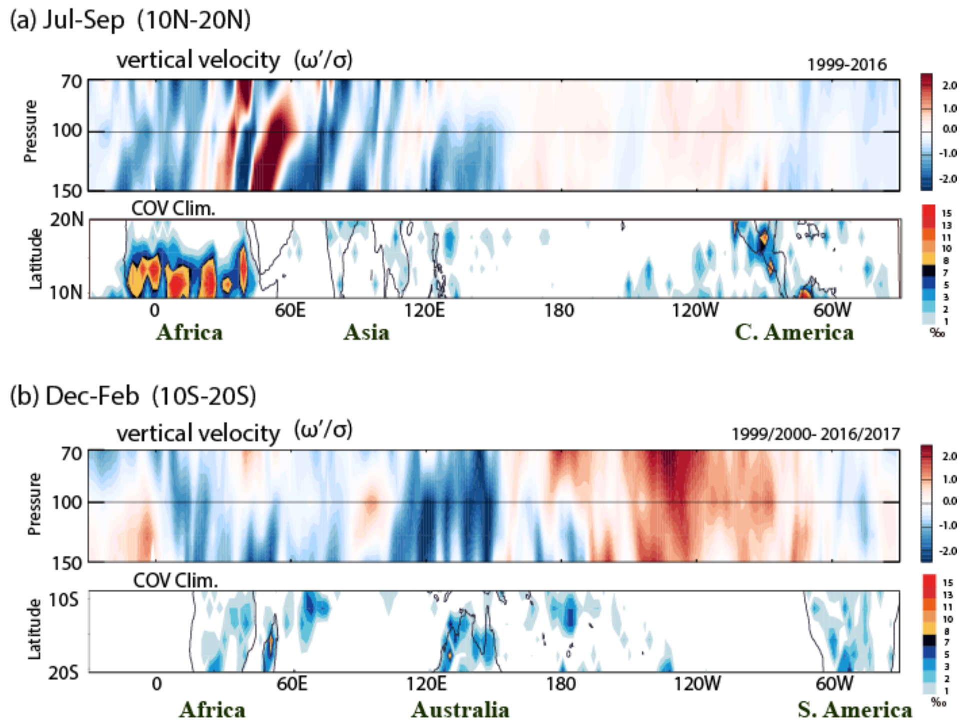

Figure 9: (a) (top panel) Height–longitude section of the standardized (with respect to the interannual variation) anomalous pressure vertical velocity averaged over 10° N–20° N, during boreal summer (JAS) 1999–2016. (bottom panel) Climatological (2007–20017) occurrence frequency of the convective over shooting (COV) in the same latitudinal zone. Unit is parts per thousand. (b) As in (a), but for 10° S–20° S in austral summer (DJF).

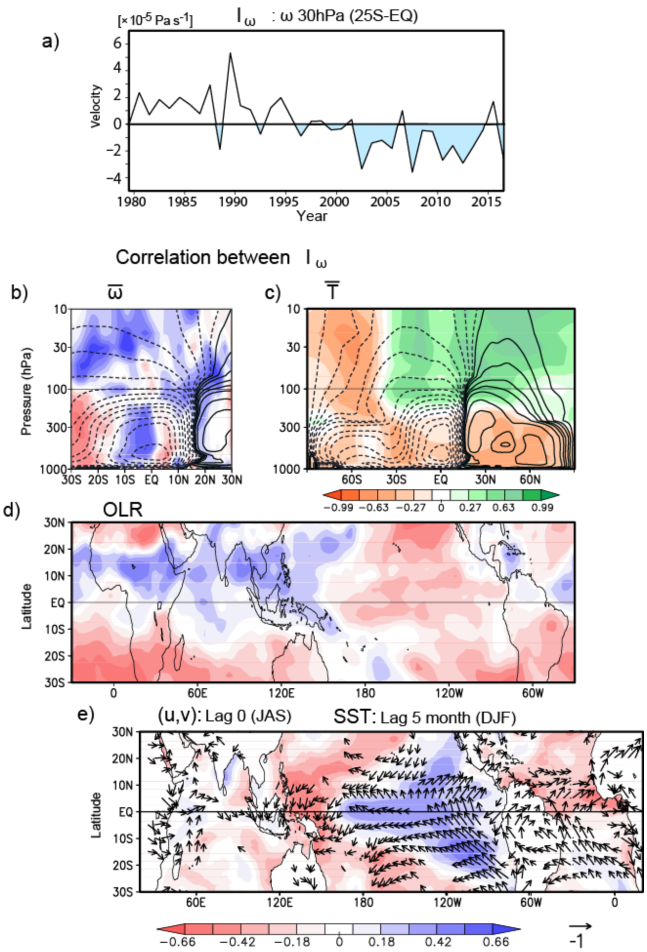

Figure 10: (a) Time series of JAS mean vertical pressure velocity (ω) at 30 hPa averaged over 0°–25° S as an index for the tropical stratospheric vertical velocity (I_ω). Correlation coefficients between I_ω and (b) zonal mean ω, (c) zonal mean temperature T at each grid, (d) OLR, and (e) horizontal winds at 925 hPa (arrows). Lagged correlation with DJF mean SST is also presented as colored shading in (e). Contours in (b) and (c) indicate the climatological residual mean meridional circulation in JAS. Solid and dashed lines indicate clockwise and counterclockwise directions, respectively.

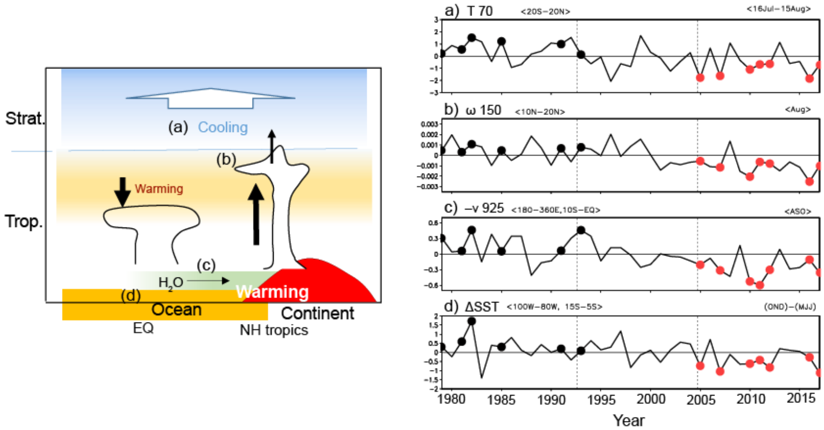

Figure 11: (left) Schematic of the recent changes in the tropics (see text). (a-c) indicate the location of the variable shown in the right panels. (Right) Time series of 4 key variables as departure from the climatology: (a) lower stratospheric temperature, (b) upwelling in the TTL, (c) cross-equatorial winds near surface, (d) SST tendency from summer to autumn. Black and red dots indicate the cases when 4 variables are same polarity–positive or negative polarity, respectively.

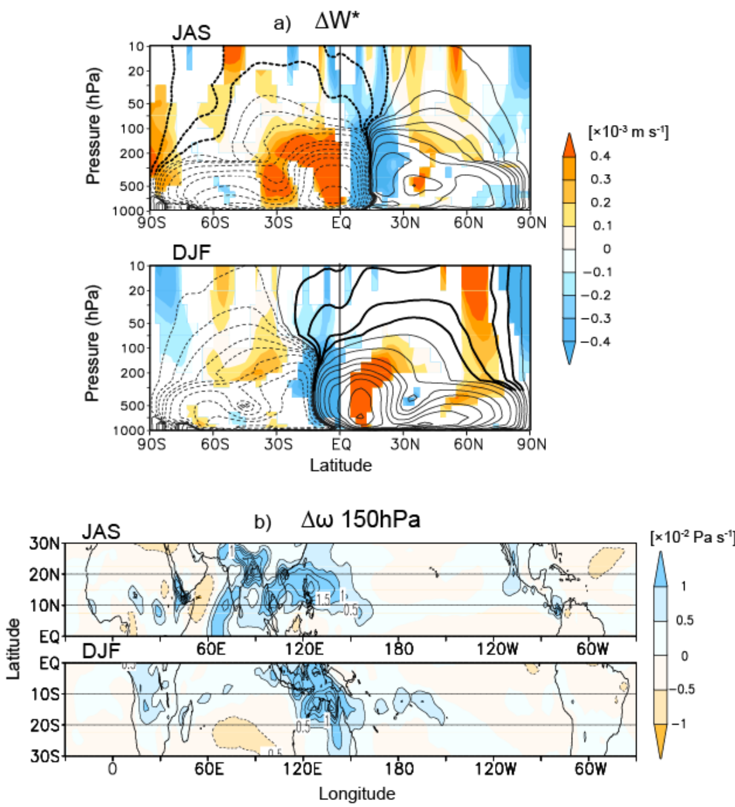

Figure 12: (a) Difference in the climatological JAS (top) and DJF (bottom) mean residual vertical velocity between JRA55-AMIP and JRA55 reanalyses (color shading). Contours indicate the mass stream function of the mean residual circulation in JRA55-AMIP. Streamlines connected to the winter polar stratosphere from the tropics are highlighted using thick lines. Solid and dashed lines represent clockwise and counterclockwise circulations, respectively. Statistically significant differences are indicated in color. (b) Difference in pressure vertical velocity at 150 hPa between JRA55-AMIP and JRA55 reanalyses in the summer tropics: (top) JAS in the NH and (bottom) DJF in the SH.