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The role of convective overshooting clouds in tropical stratosphere–troposphere dynamical coupling

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Abstract

This paper investigates the role of deep convection and overshooting convective clouds in stratosphere–troposphere dynamical coupling in the tropics during two large major stratospheric sudden warming events in January 2009 and January 2010. During both events, convective activity and precipitation increased in the equatorial Southern Hemisphere as a result of a strengthening of the Brewer–Dobson circulation induced by enhanced stratospheric planetary wave activity. Correlation coefficients between variables related to the convective activity and the vertical velocity were calculated to identify the processes connecting stratospheric variability to the troposphere. Convective overshooting clouds showed a direct relationship to lower stratospheric upwelling at around 70–50 hPa. As the tropospheric circulation change lags behind that of the stratosphere, outgoing longwave radiation shows almost no simultaneous correlation with the stratospheric upwelling. This result suggests that the stratospheric circulation change first penetrates into the troposphere through the modulation of deep convective activity.

1 **1 Introduction**

2 Weather forecasting in tropical regions is challenging due to the unstable nature of the
3 atmosphere there and its sensitivity to various extratropical disturbances. The impact of the
4 extratropical circulation on the tropics, such as the lateral propagation of tropospheric Rossby
5 waves, has been studied previously (e.g., Kiladis and Weickmann, 1992; Funatsu and Waugh,
6 2008). The influence from above (i.e., from the stratosphere) is generally neglected, but under
7 certain circumstances, such as during a sudden stratospheric warming (SSW) event,
8 stratospheric meridional circulation change can modify convective activity as will be shown
9 later.

10 Early satellite measurements showed that enhanced poleward eddy heat fluxes in the
11 extratropical stratosphere induce tropical cooling through changes in the mean meridional
12 circulation (Fritz and Soules, 1970; Plumb and Eluszkiewicz, 1999; Randel et al., 2002). It is
13 generally believed that such changes in the stratosphere do not affect the troposphere, due to
14 the difference in air density between the two. Indeed, tropical temperature change induced by
15 the intraseasonal mean meridional circulation is apparent only in the layer around 70 hPa and
16 above (Ueyama et al., 2013).

17 However, this does not imply that the stratospheric meridional circulation has no impact on
18 the atmosphere below the 70hPa level. A possible impact of stratospheric meridional
19 circulation on cumulus heating has been suggested by Thuburn and Craig (2000) in a
20 simplified general circulation model experiment. Stratospheric upwelling effects on tropical
21 convection is also confirmed by a more realistic general circulation model forecast study
22 (Kodera et al., 2011a). These models make use of cumulus parameterization to account for the
23 effect of convection into large scale circulation. Therefore, model sensitivity should be
24 dependent on the parameterization used.

25 Stratospheric effect on tropical convection is also found in non-hydrostatic models that treat
26 the convection explicitly. Although it is not fully understood yet how stability near-tropopause
27 influences anvil cloud-top height, Chae and Sherwood (2010) showed with observational data
28 and a regional non-hydrostatic model experiment that the variation of static stability near the
29 tropopause due to a change in the stratospheric upwelling, influences cloud height even if the
30 cloud height peaks only near 12 km (or 200hPa). Using a global non-hydrostatic model
31 simulation, Eguchi et al. (2014) also found that increased tropical upwelling due to a SSW

1 event reduces the static stability in the upper Tropical Tropopause layer (TTL), which leads to
2 an increase of deep convective activity in the troposphere.

3 Temperature response to stratospheric upwelling becomes unclear in the region lower than the
4 tropopause because clouds form in response to adiabatic cooling associated with upwelling.
5 Stratospheric temperature decreases, but minimal temperature changes occur in the TTL,
6 results in a decrease in static stability in the upper TTL (Li and Thompson, 2013). In the
7 regions where deep convective clouds are frequent, stratospheric influence further penetrates
8 deeper in the troposphere (Eguchi and Kodera, 2010; Kodera et al., 2011b). Once the
9 distribution of convective clouds is modified, this effect can be amplified within the
10 troposphere through a feedback involving water vapour transport (Eguchi and Kodera, 2007).

11 In a previous study composite analysis of the tropical tropospheric impact of SSW events
12 were made for the winters from 1979 to 2001 (Kodera, 2006). Even though significant
13 responses were found in the tropical troposphere, a problem of the composite analysis is that
14 by averaging many different events to extract a common feature, detailed structures often
15 become obscure. Therefore, case studies are made in the present paper on two exceptionally
16 large events focusing on the role of overshooting and deep convective clouds in stratosphere–
17 troposphere dynamical coupling in the tropics. The selected two largest SSW events of
18 January 2009 and January 2010 (Harada et al., 2010; Ayarzagüena et al., 2011) have large
19 impact on the tropical upwelling in the lower stratosphere as will be shown later. These SSWs
20 are not only large, but also localized in time unlike other SSWs. Large and simple structure of
21 the temporal variation of the forcing (eddy heat flux) and the response (stratospheric zonal
22 wind) of 2009 and 2010 SSWs permit us to investigate a detailed feature of the circulation
23 change. It should also be noted that not all major SSW events necessarily have such large
24 tropical impacts, as this depends on the latitude of the associated planetary wave breaking
25 (Taguchi, 2011).

26

27 **2 Data**

28 Meteorological reanalysis data from the European Centre for Medium-Range Forecasts
29 (ECMWF) ERA interim (Dee et al., 2011) were used to analyse air temperature and winds
30 including vertical velocity. Cloud data in the TTL, the Level 2 Cloud Layer Product
31 (Version3-01) were obtained by Cloud-Aerosol Lidar with Orthogonal Polarization
32 (CALIOP) aboard CALIPSO satellite (Winker et al., 2007). Outgoing longwave radiation

1 (OLR) data provided by NOAA (e.g., Arkin and Ardanuy, 1989) is widely used to analyse
2 convective activity in the tropics. In this study, in addition to the OLR data with a $2.5^\circ \times 2.5^\circ$
3 lat/lon resolution, we used the Microwave Humidity Sensor (MHS) channels 3 to 5 to detect
4 deep convection and convective overshoots because of the scattering by icy particles in such
5 cold precipitating clouds that causes a depression in the brightness temperatures. MHS data
6 are obtained from NOAA18 and MetOp-A. The equatorial crossing time for these platforms is
7 approximately 14h00 local time (LT) for NOAA18, and 21h30 LT for MetOp-A. In the
8 present work, the original data was regridded to a regular grid with resolution of $0.25 \text{ lat} \times$
9 0.25 lon . The figures show DC and COV occurrences resampled to a grid of 2.25×2.25 for
10 plotting purposes.

11 To capture deep, precipitating clouds we used the diagnostics developed for the tropics by
12 Hong et al. (2005), which is based on the brightness temperature differences (ΔT) measured
13 by three channels of the MHS between: i) 183.3 ± 1 and 183.3 ± 7 GHz (ΔT_{17}); ii) 183.3 ± 1
14 and 183.3 ± 3 GHz (ΔT_{13}); and iii) 183.3 ± 3 and 183.3 ± 7 GHz (ΔT_{37}). Deep convective
15 cloud (DC) and convective overshooting (COV) were discriminated according to the
16 following criteria, in which COV refers to clouds able to penetrate into the tropopause region
17 (Hong et al., 2005; Funatsu et al., 2012). Deep convective cloud: $\Delta T_{17} \geq 0$, $\Delta T_{13} \geq 0$, ΔT_{37}
18 ≥ 0 K; and convective overshooting: $\Delta T_{17} \geq \Delta T_{13} \geq \Delta T_{37} > 0$ K.

19 Although these high frequencies are generally not sensitive to cirrus and anvil cirrus clouds,
20 they will probably have difficulty distinguishing some strong anvil clouds from deep
21 convective clouds. But fortunately, these strong anvil clouds are generally tightly connected
22 with deep convective cloud systems (Hong et al., 2008).

23 The Tropical Rainfall Measuring Mission (TRMM) daily-integrated precipitation (TRMM
24 3B42 v7) was used to study surface precipitation (Huffman et al., 2007).

25

26 **3 Results**

27 An enhanced Brewer-Dobson (BD) circulation during a stratospheric warming event creates
28 strong downwelling in the polar region and upwelling in the tropical stratosphere, and thus
29 warming and cooling tendency in these respective regions. Figures 1a and 1b show the
30 evolution of eddy heat flux at 100 hPa averaged over the extratropical Northern Hemisphere
31 (NH; 45°N – 75°N), and the latitude–time section of the zonal mean pressure coordinate

1 vertical velocity at 50 hPa from 1 January to 11 February (the left and right panels are for
 2 2009 and 2010, respectively). In both years, stratospheric upwelling in the tropics at the 50
 3 hPa level strengthens following the increase in wave activity at around 16 January in 2009,
 4 and around 20 January 2010 (indicated by the solid vertical lines in the figure). In the tropics,
 5 an increase in COV is synchronous with the stratospheric upwelling (Fig. 1c). The convective
 6 activity represented by the OLR also increases in the Southern Hemisphere (SH), which can
 7 also be characterized as a southward shift of the active convective region (Fig. 1d). A delay in
 8 the response of the OLR in the SH is also noted. The difference in the characteristics in the
 9 temporal variation in COV and OLR relative to the vertical velocity at 50 hPa becomes also
 10 apparent in the vertical structure of the correlation coefficient in the following.

11 To study the relationship between tropospheric convective activity and the vertical velocity at
 12 different pressure levels, correlation coefficients were calculated between variables
 13 representing a convective activity (COV, DC, and OLR) and the pressure vertical velocity (ω)
 14 at each level (Fig. 2). **These correlation coefficients are simply being used to identify the**
 15 **relation between dynamical variables in two rather short-duration events.** Variables were first
 16 averaged over the tropics (25°S to 25°N) and then correlations were calculated for the 31 day
 17 period centred on the onset day (16 January for 2009 and 20 January for 2010). For
 18 convenience of comparison, the sign of the OLR was reversed ($-\text{OLR}$). In both winters, COV
 19 shows the highest correlation with ω in the lower stratosphere around 70–50 hPa. DC is also
 20 correlated with the stratospheric upwelling, but less so. The OLR shows little relationship
 21 with the stratospheric circulation, although it is correlated with vertical velocity in the upper
 22 troposphere.

23 Here, we check the physical consistency among the variables by comparing the correlation
 24 coefficients among them. It is reasonable to expect that stratospheric vertical velocity should
 25 have the strongest relationship with the occurrence of COV (i.e., convection penetrating to the
 26 stratosphere) and the weakest relationship with OLR, which is sensitive to lower clouds as
 27 well as deep convection. Therefore, the following inequalities among the correlation
 28 coefficient, r , between the lower stratospheric pressure vertical velocity, ω , should be
 29 expected:

$$30 \quad r_{\omega, \text{COV}} < 0, \quad |r_{\omega, \text{COV}}| > |r_{\omega, \text{DC}}|, \quad |r_{\omega, \text{DC}}| > |r_{\omega, -\text{OLR}}|, \quad (1)$$

31 where $r_{\omega, \text{COV}}$, $r_{\omega, \text{DC}}$, and $r_{\omega, -\text{OLR}}$ are the correlation coefficients between ω and COV, DC, or –
 32 OLR, respectively.

1 Such relationship is satisfied in the correlation analysis presented in Fig. 2. This result
2 supports our working hypothesis that lower stratospheric vertical velocity variation is coupled
3 with the tropical convective activity.

4 The present study can also be compared with a regression study of BD circulation index by Li
5 and Thompson (2013); Enhanced BD circulation increases clouds occurrence above the
6 tropical tropopause, in association with a decrease of stratospheric temperature and the static
7 stability around the tropopause. The structure of the tropical temperature and stability change
8 associated with the COV is consistent with a variation associated with a strengthening of the
9 BD circulation. Formation of the clouds above the tropopause is also consistent with the
10 correlation of COV with upwelling above 100 hPa.

11 Figure 3 depicts a development of downward coupling in the equatorial summer tropics,
12 averaged between 20°S and the equator. The temperature tendency (Fig. 3a) shows a rapid
13 decrease in the stratosphere following the increase in the eddy heat flux in Fig. 2a, but no
14 clear temperature signal is observed in the troposphere, which agrees with the results of
15 previous study (Ueyama et al., 2013). Figure 3b shows altitude-time section of measured
16 cloud frequency (optical thickness < 4) by CALIOP. Horizontal dashed lines indicate
17 approximate height corresponding to 100 hPa pressure level (solid lines in Fig. 3a and 3c).
18 Prior to the SSWs, thin clouds are formed near 16.6 km (or 100 hPa) around a cold point
19 tropopause. When cooling events start, cloud forms all the depth of the TTL, indicating a
20 development of convective activity. Pressure vertical velocity is shown as departure from the
21 period mean normalized by a daily standard deviation at each level to visualize the large range
22 of variation (Fig. 3c). Although vertical velocity varies in a similar manner to temperature
23 tendency in the stratosphere, an increase in the upwelling also occurs in the troposphere
24 following the stratospheric change. This tropospheric upwelling is associated with an increase
25 in surface precipitation (Fig. 3d).

26 This result shows that the temperature tendency is a good proxy for vertical velocity in the
27 stratosphere. However, dynamical cooling tends to be compensated by diabatic heating due to
28 cloud formation lower than the tropopause as illustrated in Fig. 3; consequently, the
29 temperature tendency is no longer a good indicator of the vertical velocity below 70 hPa.

30 Figure 4 shows the evolution of the geographical distribution of OLR and COV before (i), and
31 after (ii) the onset of the event. The influence of the El Niño Southern Oscillation (ENSO) is
32 evident in the OLR during period (i). In January 2009, which is a cold phase of ENSO, a well-

1 developed region of low OLR is located over the Maritime Continent, while in January 2010,
2 a warm phase of ENSO, it is located over the western Pacific according to the change in the
3 equatorial Pacific sea surface temperature (SST). The velocity potential at 925 hPa (contour
4 lines) in period (i) indicates that these convective activities are maintained by a large-scale
5 low-level convergence. After the onset of the stratospheric event during period (ii), the low-
6 OLR centre over the Maritime Continent or western Pacific is weakened, and multiple
7 convective-active regions develop in the SH along 15°S. This active convective zone includes
8 tropical cyclones and storms (names are indicated below the panel) over warm ocean sectors
9 near Madagascar, North of Australia, and in the southwestern Pacific.

10 The occurrence of COV is high over the African and South American continents, but no
11 particular enhancement is seen around the Maritime Continent–western Pacific region in
12 period (i). This indicates the weaker dependency of COV on low-level convergence. Although
13 the occurrence of COV increases after the onset in period (ii), no substantial change is seen in
14 the spatial structure except that the COV distribution takes a more zonal form. The
15 distribution of the regions with low OLR becomes increasingly similar to that of COV during
16 period (ii). This indicates that the COV-related deep convective activity becomes important
17 after the onset of the stratospheric event.

18

19 **4 Summary and discussion**

20 The results of our analysis of changes in tropical circulation associated with large SSWs
21 during January 2009 and January 2010 can be summarized as follows.

22 Enhanced stratospheric wave activity produced a cooling in the tropical stratosphere through a
23 strengthening of the BD circulation. This influence penetrated downward into the troposphere
24 through a change in the cloud formation. Among the variables representing different
25 convective activity, COV shows the highest correlation with the lower stratospheric vertical
26 velocity. This result is reasonable because the COV clouds penetrate above the tropopause
27 and interact directly with the stratospheric circulation. The reason of low correlation of the
28 OLR with stratospheric upwelling originates from the fact that the tropospheric variation lags
29 by about a week (Fig. 1).

30 The results obtained from the present two SSW events are consistent with the earlier results
31 from an independent composite analysis of the NH winters for a period of 1979 to 2001.

1 Figure 5a shows the results of the above mentioned composite analysis. Twelve SSW events
2 of which maximum deceleration of the polar night jet (average 50°N-70°N) at 10hPa exceeds
3 $2\text{ms}^{-1}/\text{day}$ with a smoothed data are selected (see detail in Kodera 2006). The key day is
4 defined as the day of the largest deceleration. Student- t values corresponding to a 95%
5 significance level for one- and two-sided tests are 1.8 and 2.2, respectively. Following a
6 deceleration of the polar night jet, statistically significant increase in the upwelling occurs in
7 the tropical stratosphere around day 2, and in the tropospheric equatorial SH around day 4 to
8 11.

9 Two SSW events in the present study are juxtaposed below in Fig. 5b. The top panel shows
10 the zonal-mean zonal wind tendency of winters 2009 and 2010 similar to Fig 5a-top panel.
11 The tropical vertical pressure velocity in the SH (20°S-Eq) is presented in a similar way as the
12 composite analysis by choosing the day of the maximum deceleration as the time origin. We
13 can see that the upwelling in the tropical SH increases in the upper troposphere around day 4
14 to day 11 similarly to the composite mean (rectangles in Fig. 5). **Therefore the relationship**
15 **that we have identified here in two particularly strong SSWs between the SSW and the**
16 **enhancement of tropical convection is consistent with that previously identified from the**
17 **composite analysis.**

18 To get an insight into a possible mechanism of connection between the stratospheric and
19 tropospheric variability, we also calculated correlations between the temperature or vertical
20 temperature gradient (or static stability) at each level, and COV or -OLR (Fig. 2 bottom).
21 COV shows stronger relationship around the tropopause with vertical temperature gradient
22 (Fig. 2e) than temperature itself (Fig. 2d). This means that COV is sensitive to the stability
23 around the tropopause region (100 hPa), while OLR is related with the static stability in the
24 upper troposphere (Fig. 2f). This result indicates that COV increases due to a decrease of
25 static stability around the tropopause induced by a cooling in the lower stratosphere
26 associated with the SSW, consistent with the results of Kuang and Bretherton (2004) and
27 Chae and Sherwood (2010). Our previous numerical experiment also shows that when local
28 cooling occurs near the tropopause, upwelling enhances accompanying a warming in the
29 lower TTL and the upper troposphere (see Figure 4 of Kodera et al., 2011a). A global non-
30 hydrostatic model study (Eguchi et al., 2014) also confirmed the relationship suggested in the
31 present result. Therefore, we consider that although the cooling effect by stratospheric

1 upwelling is limited in the stratosphere, its effect can further penetrate below through changes
2 in COV and deep convective activity.

3 Changes were also noted in the spatial distribution of the convective activity following the
4 stratospheric event (Figure 4). When stratospheric upwelling was suppressed before the onset
5 of the event (period i), convection tended to cluster around the equatorial Maritime Continent
6 or western Pacific region depending on the phase of ENSO. When the stratospheric upwelling
7 increased (period ii), convection expanded over a wide range of longitudes in the tropical
8 summer hemisphere. In other words, tropical circulation changed from a more Walker like
9 (east–west) configuration to a more Hadley (north–south) type.

10 The Madden–Julian Oscillation (MJO) (Madden and Julian, 1994) has a significant influence
11 on tropical convective activity. It is reported that the occurrence of the SSW is related with
12 the phase of the MJO (Garfinkel et al, 2012; Liu et al 2014). One would ask whether or not
13 the present phenomenon is associated with the MJO. The features of the MJO in January 2009
14 and 2010 differed significantly as can be seen in Figure 6. A convective centre remained
15 stationary over the Maritime Continent prior to the onset of the 2009 stratospheric event, after
16 which an eastward propagation was initiated from the Indian Ocean. In contrast, an eastward
17 propagating convective centre became almost stationary over the western Pacific after the
18 onset in January 2010. In spite of the differences in the MJO in January 2009 and 2010,
19 circulation changes related to the stratospheric events showed similar features during both
20 winters, suggesting that the present phenomenon is independent of the MJO.

21 **The certainty of the dynamical connections identified here is of course limited by the small
22 number and the relatively short duration of the events. Further certainty will come from future
23 modelling studies and observational studies of a larger set of events.**

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31 were from Atmospheric Science Data Center (ASDC) at NASA. TRMM data were acquired
32 through the Giovanni online data system, developed and maintained by NASA GES DISC.

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- 15

1 Figure captions

2 Figure 1. a) Time series of the eddy heat flux at 100 hPa averaged over 45°N–75°N [K ms^{-1}].
3 b) Zonal mean pressure coordinate vertical velocity at 50 hPa [Pa s^{-1}]. c) Number of
4 convective overshootings per day at each latitude. d) Zonal mean OLR [W m^{-2}]. Variables are
5 displayed from 1 January to 11 February. Left- and right-hand panels are for 2009 and 2010,
6 respectively. Vertical velocity and OLR data are smoothed by a three-day running mean.

7

8 Figure 2. a) Correlation coefficient between the pressure coordinate vertical velocity (ω) at
9 each pressure level and the daily convective overshooting occurrence frequency (COV)
10 averaged over the tropics. b) As for (a), but for deep convection (DC). c) As for (a), but for
11 the correlation coefficient with $-\text{OLR}$. d) Same as in (a), except for COV and temperature at
12 each level. e) Same as in (d) except for COV and vertical temperature gradient at each level,
13 f) Same as in (e) , except for $-\text{OLR}$ and vertical temperature gradient. Variables were first
14 averaged over 25°S to 25°N and then the correlation was calculated over 31 days centered at
15 the onset day (16 January in 2009 and 20 January in 2010). Solid and dashed lines indicate
16 2009 and 2010, respectively.

17

18 Figure 3. a) Similar to Fig. 1, except for the pressure–time section of the zonal mean
19 temperature tendency averaged over the SH tropics (20°S to the equator) [K day^{-1}]. b) As for
20 (a), except for the geographical altitude-time section of cloud frequency measured by
21 CALIOP [%]. (c) As for (a), except for the pressure coordinate vertical velocity anomalies
22 normalized by the standard deviation of daily variability. d) Time series of the daily TRMM
23 surface precipitation averaged over SH tropics [mm day^{-1}]. Horizontal solid lines in (a) and
24 (c) and dashed lines in (b) indicate 100 hPa pressure level.

25

26 Figure 4. (a, c, e, g): seven-day mean OLR (color shadings) with velocity potential at 925 hPa
27 (contours of 6, and $8 \times 10^6 \text{ m}^2 \text{ s}^{-1}$). (b, d, f, h): seven-day average of the number of COV in
28 each 2.5° lat/lon grid box. (a,b) and (c,d) are seven-day period before (i) and after (ii) the
29 onset of the event in January 2009. (e,f) and (g, h) are the same as (a,b) and (c,d), except for
30 the event in January 2010.

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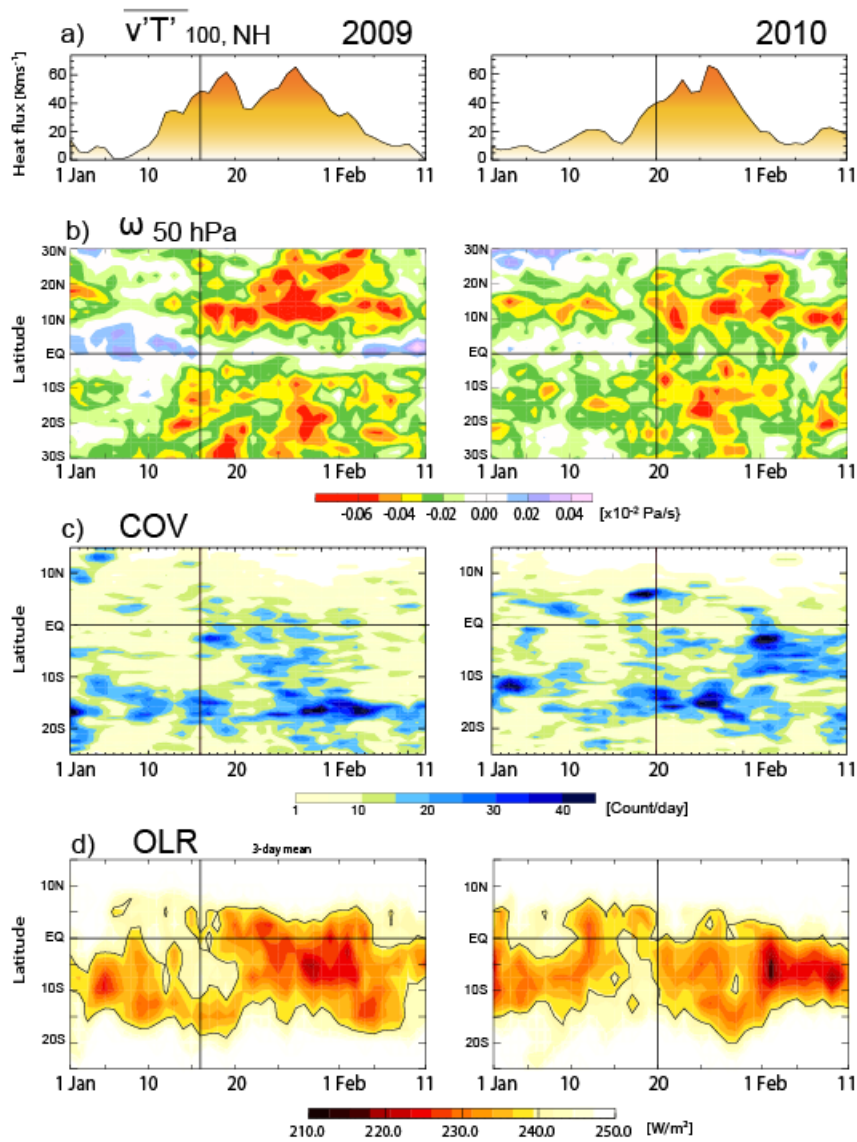
2 Figure 5 (a) Composite analysis of twelve SSWs during boreal winters from 1979- 2001 (see
3 Kodera (2006) for detail): Low pass filtered zonal-mean zonal wind tendency at 10 hPa
4 averaged over 50°-70°N of twelve events (top). Student-t values of composited vertical
5 pressure velocity averaged over 30°S-30°N in the stratosphere (middle) and that of 10°S-
6 Equator in the troposphere. (b) Zonal-mean zonal wind tendency in winters 2009 and 2010
7 similar to Figure 7a (top). Normalized tropical vertical pressure velocity averaged over 20°S-
8 Equator in January 2009 (middle) and January 2010 (bottom). Vertical lines indicate key date
9 (see text). Rectangles indicate a period of enhanced tropospheric upwelling in (a).

10

11 Figure 6. Time–longitude sections of three-day running mean equatorial (5°S–5°N) OLR over
12 the Indian Ocean–central Pacific sector (30°E–150°W) during boreal winter for (left)
13 2008/2009 and (right) 2009/2010. The figure displays a two-month period centered on the
14 onset day of the tropical stratospheric upwelling events (16 January 2009 and 20 January
15 2010) indicated by horizontal solid lines.

16

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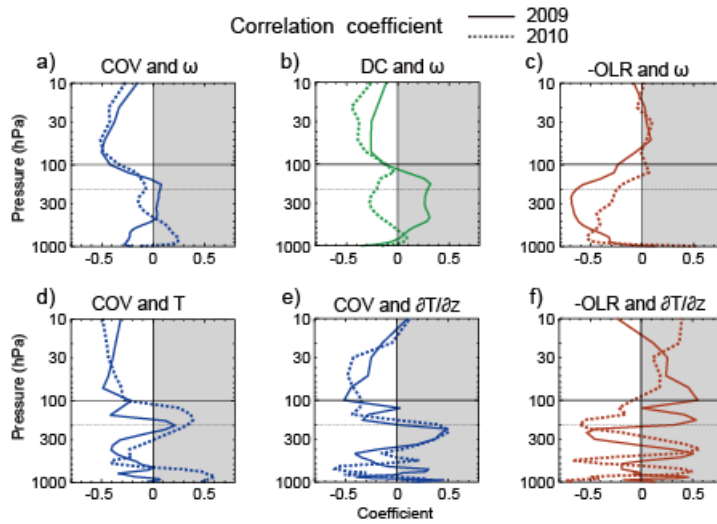


2

3 Figure 1. a) Time series of the eddy heat flux at 100 hPa averaged over 45°N–75°N [K ms^{-1}].
 4 b) Zonal mean pressure coordinate vertical velocity at 50 hPa [Pa s^{-1}]. c) Number of
 5 convective overshootings per day at each latitude. d) Zonal mean OLR [W m^{-2}]. Variables are
 6 displayed from 1 January to 11 February. Left- and right-hand panels are for 2009 and 2010,
 7 respectively. Vertical velocity and OLR data are smoothed by a three-day running mean.

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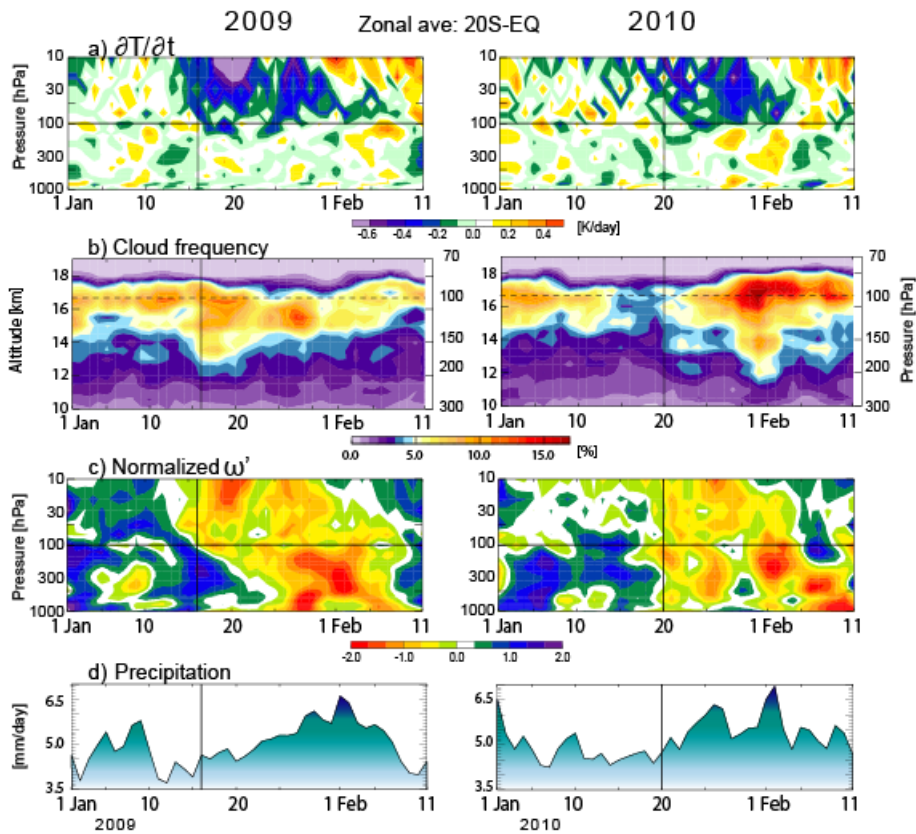


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2 Figure 2. a) Correlation coefficient between the pressure coordinate vertical velocity (ω) at
 3 each pressure level and the daily convective overshooting occurrence frequency (COV)
 4 averaged over the tropics. b) As for (a), but for deep convection (DC). c) As for (a), but for
 5 the correlation coefficient with $-OLR$. d) Same as in (a), except for COV and temperature at
 6 each level. e) Same as in (d) except for COV and vertical temperature gradient at each level,
 7 f) Same as in (e) , except for $-OLR$ and vertical temperature gradient. Variables were first
 8 averaged over $25^{\circ}S$ to $25^{\circ}N$ and then the correlation was calculated over 31 days centered at
 9 the onset day (16 January in 2009 and 20 January in 2010). Solid and dashed lines indicate
 10 2009 and 2010, respectively.

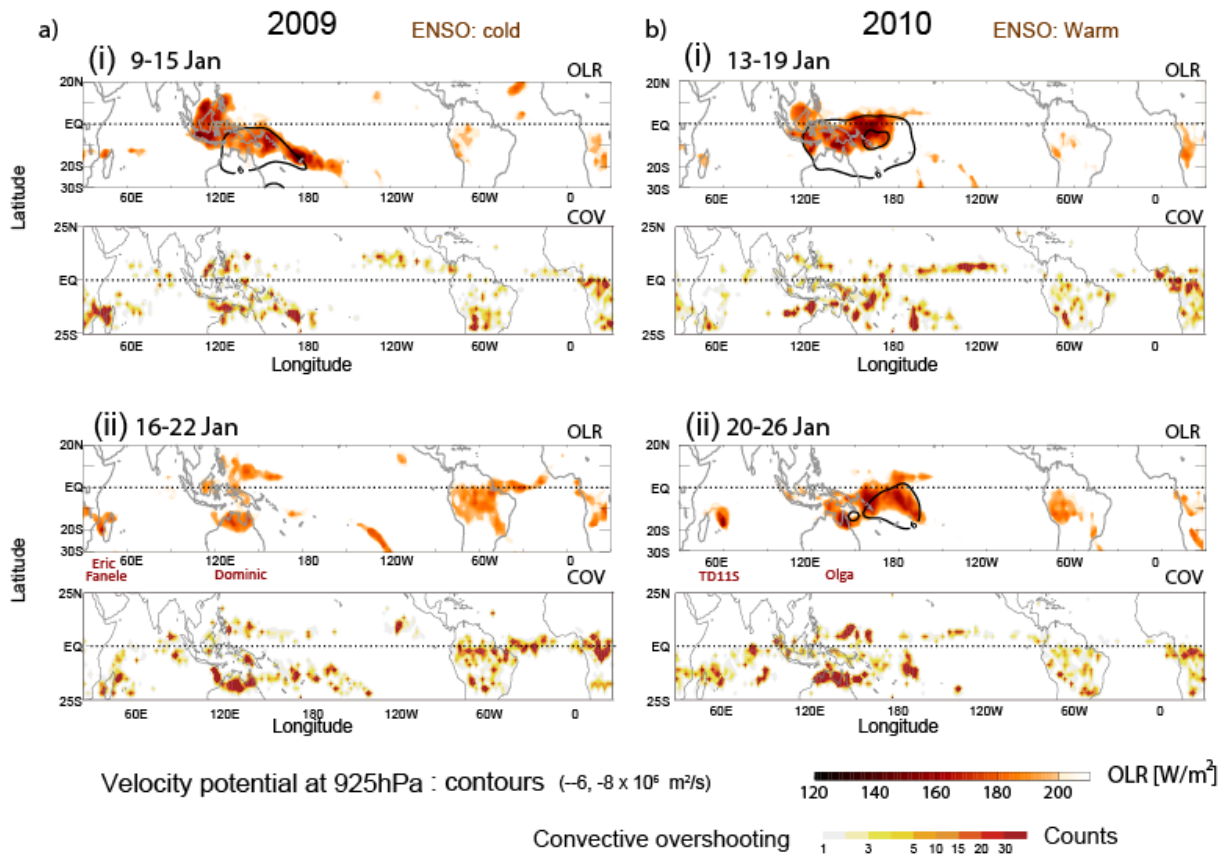
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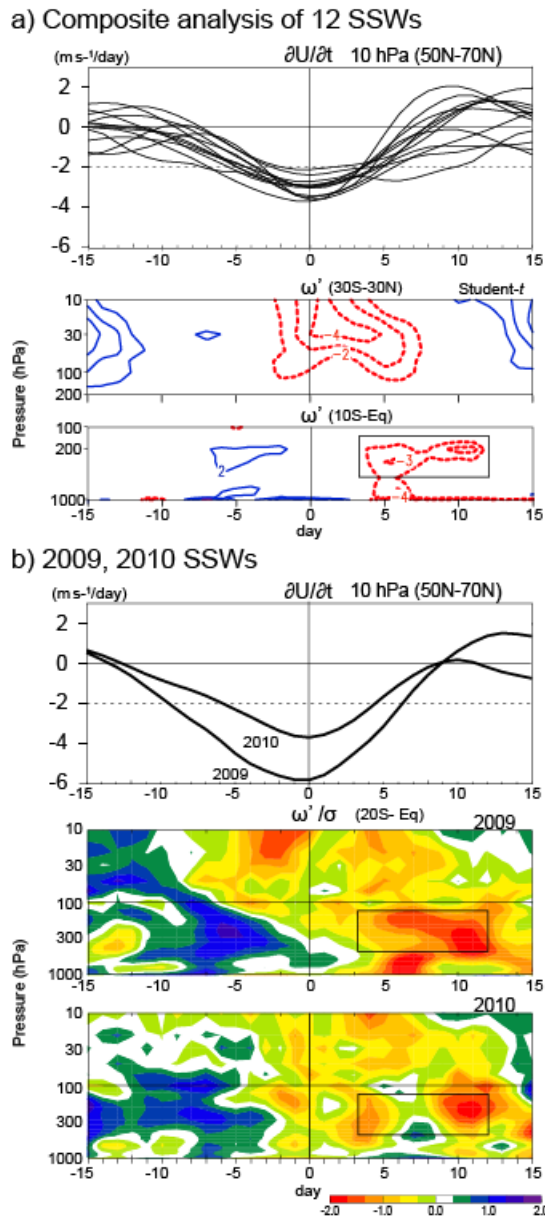
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7



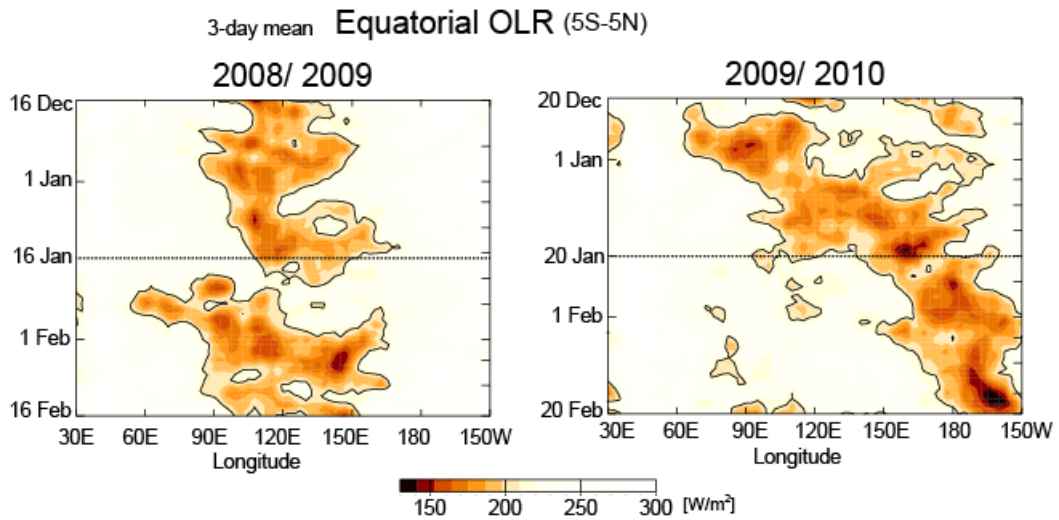
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 6 Equator in the troposphere (bottom). (b) Zonal-mean zonal wind tendency in winters 2009
 7 and 2010 similar to (a) top panel. Normalized tropical vertical pressure velocity averaged over
 8 20°S-Equator in January 2009 (middle) and January 2010 (bottom). Vertical lines indicate
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