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pdftitle=Downward coupling in the TTL  
pdfauthor=N. Eguchi et al.

title Manuscript prepared for Atmos. Chem. Phys. Discuss.  
with version 4.1 of the L<sup>A</sup>T<sub>E</sub>X class copernicus\_discussions.cls.  
Date: 3 July 2014

# **A global non-hydrostatic model study of a downward coupling through the tropical tropopause layer during a stratospheric sudden warming**

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## Abstract

The dynamical coupling process between the stratosphere and troposphere in the tropical tropopause layer (TTL) during a stratospheric sudden warming (SSW) in boreal winter was investigated using simulation data from a global non-hydrostatic model (NICAM) that does not use cumulus parameterization. The model reproduced well the observed tropical tropospheric changes during the SSW including the enhancement of convective activity following the amplification of planetary waves. Deep convective activity was enhanced in the latitude zone 20–10° S, in particular over the southwest Pacific and southwest Indian Ocean. Although the upwelling in the TTL was correlated with that in the stratosphere, the temperature tendency in the TTL was mainly controlled by diabatic heating originating from cloud formation. This result suggests that the stratospheric meridional circulation affects cloud formation in the TTL.

## 1 Introduction

There have been many studies on the interaction between the stratosphere and troposphere in the tropical tropopause layer (hereafter, TTL), which is typically located at 14–19 km altitude or 150–90 hPa (Highwood and Hoskins, 1998; Fueglistaler et al., 2009), with stratospheric water vapor variation on interannual and seasonal time scales (cf., Randel and Jensen, 2013) an important topic. Possible impacts of the recent stratospheric cooling trend on the troposphere, such as an increase in tropical cyclones, are currently the subject of discussion (Emanuel et al., 2013; Ramsay, 2013). However, it is difficult to separate the stratospheric effect on the troposphere from other factors for long-term variations such as a warming of the ocean.

Here, we focus on changes in the troposphere during a stratospheric sudden warming (hereafter, SSW) event, which drastically modifies the stratospheric circulation in the space of a week. It is, therefore, relatively easy to identify the causal relationship between the stratospheric and tropospheric variation, which can be separated from long-term variability such as the Madden-Julian Oscillation (MJO), El Niño–Southern Oscillation (ENSO), or the Quasi-Biennial Oscillation (QBO). Previous observational studies showed that strong SSW events modulated tropical

convective activity and the general circulation in the troposphere (the Hadley and Walker cells) (Eguchi and Kodera, 2007, 2010 (EK10)). However, it is still not clear how the connection between the stratosphere and troposphere occurs and how it modulates the convective activity during SSW events.

5 This is largely due to a lack of global observations of vertical velocity in the TTL. Ueyama et al. (2013) showed that the temperature tendency in the stratosphere can be used as a proxy for the vertical motion (i.e., the strength of the Brewer–Dobson (BD) circulation). However, this relationship does not hold below the tropopause, where diabatic heating due to cloud formation cancels adiabatic cooling, as will be shown later.

10 Data from numerical simulation are useful for examining vertical velocity in the TTL. Thuburn and Craig (2000) used momentum forcing in a simplified global general circulation model (GCM) to show that a change in the stratospheric meridional circulation modified cumulus heating in the TTL as well as the tropopause height. Kodera et al. (2011) constructed a more realistic forecast experiment by incorporating an atmospheric blocking-type circulation anomaly in the  
15 North Atlantic in the initial conditions to amplify planetary waves and produce strong BD circulation later in the stratosphere. The results showed similar effects in the tropics to those seen in observational studies, that is, the tropical convective activity was enhanced zonally, especially in the Southern Hemisphere (SH), and cooling at the tropopause region associated with the SSW event was capable of modulating tropical convective activity.

20 These results suggest that convection plays an important role in the stratosphere–troposphere coupling in the tropics. In these models, however, the effects of convection were not explicitly treated, but were represented by cumulus parameterization. Further, to investigate the adiabatic and diabatic parameters in the TTL is necessary to clarify the coupling process between the stratosphere and troposphere through the thermo-dynamic balance. In the present study,  
25 we investigate changes in convection during a SSW event that occurred in January 2010 using simulation data from a global non-hydrostatic model, NICAM (non-hydrostatic icosahedral atmospheric model) (Satoh et al., 2008), which does not use cumulus parameterization. The advantage using the NICAM data is able to discuss the dynamic and thermodynamic changes in the TTL due to the changes of BD circulation and convection around SSW event. A case

study using such high-resolution simulation data has the advantage of capturing the fine vertical and temporal structures of the TTL during the SSW event, which might be smoothed away in a statistical analysis of a huge number of events occurring under a range of conditions.

The remainder of the paper is organized as follows. The next section deals with the NICAM simulation data (Sect. 2), Sect. 3 shows the SSW event simulated in NICAM and the dynamical variability in the stratosphere and troposphere during the SSW. A summary and discussion of the results are given in Sect. 4.

## 2 Analysis data

The numerical simulation was conducted using NICAM, the global non-hydrostatic model with a horizontal mesh size of 14 km and 40 vertical levels. The model top was at 38 km with nine layers between 10 km and 20 km altitude. Moist convection was explicitly calculated using a cloud microphysical scheme (single moment scheme) with six prognostic variables (Tomita, 2008). Atmospheric radiation, turbulence, and ocean processes were calculated using the MSTRN-X (Nakajima et al., 2000; Sekiguchi and Nakajima, 2008), Mellor–Yamada–Nakanishi–Niino level 2 (Mellor and Yamada, 1982; Nakanishi and Niino, 2006; Noda et al., 2010), and slab ocean models, respectively. The sea surface temperature (SST) was initialized using the National Centers for Environmental Prediction (NCEP) final (FNL) operational global analysis data ( $1^\circ \times 1^\circ$ ), and nudged to the NOAA weekly Optimum Interpolation SST analysis (Reynolds and Smith, 1994) with a relaxation time of seven days.

The atmospheric initial data were interpolated from the ECMWF YOTC (Year of tropical Convection) analysis (Waliser et al., 2012). The simulation period was the 60 days starting from 20 December 2009. During this period, a significant MJO event took place (Waliser et al., 2012), and this MJO event was targeted in a model intercomparison project (<http://yotc.ucar.edu/mjo/vertical-structure-and-diabatic-processes-mjo>). The cloud properties in the upper troposphere (UT) have been evaluated in comparison with satellite observations (e.g., Inoue et al., 2010; Kodama et al., 2012).

In the present study, winds (zonal, meridional and vertical), temperature, specific humidity, diabatic heating rates (by cloud microphysics and solar radiation), snow, ice and graupel contents, and column integrated cloud fraction were analyzed. The cloud fraction data are two-dimensional (longitude–latitude), and the others are three-dimensional. Snapshot data were archived at 3 h intervals except for cloud fraction (hourly mean) and diabatic heating rate (daily mean). All the output variables were daily averaged and converted into  $1^\circ$  datasets.

## 3 Results

### 3.1 Stratospheric sudden warming simulated by NICAM

In the real atmosphere, the SSW in the boreal hemisphere occurred during January 2010 (cf., Dörnbrack et al., 2012). The temperature increased by  $+35$  K in the northern polar region and decreased by  $-2.5$  K in the tropical lower stratosphere (LS) with increasing wave activity at middle latitudes (Fig. 1a and b). In the NICAM simulation, minor stratospheric warming spontaneously occurred in January 2010, although the date on which the tropical lower stratospheric (approximately 50–80 hPa) temperature started decreasing was approximately five days earlier than in the real atmosphere. The temperature rose in the polar region, but decreased in the tropics (Fig. 2a) because enhanced wave activity at middle latitudes (Fig. 4c) induced downwelling in the polar region and upwelling in the tropics in the stratosphere. The temperature variation was comparable to that in the real atmosphere: the temperature rose by approximately  $+25$  K in the northern polar region and fell  $-1.5$  K in the tropics after 7 January.

The tropical averaged temperature in the LS decreased gradually from 14 January; the latitude–time section of the LS temperature (Fig. 2b) shows that cooling on the south side of the equator starts around 14 January before that on the north side. In the tropical troposphere, the center of the region of active convection shifted southward from  $5^\circ$  S to  $15^\circ$  S after 14 January, when the LS started cooling. The latitude of minimum temperature was located to the south of the convective region (around  $25^\circ$  S; Fig. 2c), here the LS upwelling was enhanced in period (ii).

After 21 January, cooling occurred over the wide tropics ( $30^{\circ}\text{S}$ – $30^{\circ}\text{N}$ ) with an equatorially symmetric structure (Fig. 2b). The largest column integrated cloud fraction occurred in association with the cooling in the tropical LS around 24 January. Note that the column integrated cloud fraction includes both convective and upper level thin clouds. The convective clouds are measured by outgoing longwave radiation (OLR). The region of minimum OLR shifted southward from  $5^{\circ}\text{S}$  to  $12.5^{\circ}\text{S}$  after 14 January, and the upper level clouds (ice clouds) extended farther southward ( $15^{\circ}\text{S}$ ) than convective clouds (Fig. 2d). In the real atmosphere, the convective region was shifted southward around 20 January and enhanced around 2 February (Fig. 1c). The convective initiation of the MJO event took place over the western Indian Ocean at the end of December 2009, and the MJO propagated eastward during January 2010. The SSW occurred when the MJO passed over the maritime continent. In the simulation, the MJO was weaker than in the observations, the eastward propagation started at the end of period (ii) or the beginning of the period (iii) (figure is not shown). Further, the equatorial convection in the simulation (also observation) was suppressed during the period (ii) (when convection was passing over the maritime continent in the observation), because the convective activity regions shifted southward off equatorial southern hemisphere. It is suggested that the convection behavior cannot be explained only by eastward travelling convective signal, such as MJO. Thus, the effects of the MJO are expected to be small for the simulated convective variations during the target two weeks.

We divided the period into three consecutive seven-day periods according to the LS tropical temperature (Fig. 2): period (i) (7–13 January) is prior to the start of the cooling event. Period (ii) (14–20 January) is characterized by a cooling trend, while period (iii) (21–27 January) is the period of peak cooling. These periods could be regarded as the initiation, transition, and mature phases of the SSW impact on the tropics, respectively.

### 3.2 Connection between stratospheric and tropospheric meridional circulation

Ueyama et al. (2013) demonstrated that the tropical temperature tendency at 70 hPa (LS) is closely related to the strength of the stratospheric meridional circulation. To investigate in more detail the structure of the coupling between the stratospheric and tropospheric circulation, the

temporal correlation coefficient between the tropical mean vertical velocity at 70 hPa and the zonal mean vertical velocity at each level was calculated (Fig. 3a). The vertical velocity at 100 hPa (TTL) was not correlated with that at 70 hPa, except near 20° S. However, better correlation is found in the deep troposphere and the LS in a latitudinal zone of 20° S–10° S where the convection is enhanced in period (ii) as shown in Fig. 2c. These features are also confirmed in time-longitude sections (not shown).

Variation of the vertical velocity in the troposphere usually accompanies a change in cloud formation. The difference in diabatic heating due to cloud microphysics between periods (ii) and (i) is displayed in Fig. 3b. A large change in the diabatic heating is found between 20–10° S corresponding to the zone of enhanced upwelling from the surface to the TTL. In the TTL (at 100 hPa), an increase of the diabatic heating extended all over the tropics including the Northern Hemisphere (NH). This zone of increased diabatic heating just below 70 hPa corresponds to a zone of decreased water vapor (Fig. 3c), as upwelling induces adiabatic cooling leading to sublimation of water vapor. The cooling trend in the TTL was evident over the wide tropics (Fig. 2b).

Cirrus cloud forms in the TTL (30° S–20° N) after the onset of the SSW, although higher correlations of vertical wind with that over the tropical LS are limited to within the convective region (20–10° S). In other words, the low correlation between the vertical motion in the LS and that in the TTL (around 100 hPa) implies that the mechanisms of the formation of ice clouds in the TTL were different from those of convective clouds in the tropical deep troposphere. The next subsection will describe the detailed mechanism of ice cloud formation in the TTL.

Figure 3b and c also shows that the increased diabatic heating in the deep troposphere occurs in association with an increase of water vapor in the lower troposphere. This is indicative of positive feedback between convective activity and low-level moisture convergence in the lower troposphere (EK10) over the tropical SH in period (iii) as illustrated in Fig. 2c.

### 3.3 Downward propagation of dynamical signal through the TTL

Time–altitude sections of the temperature tendency, static stability and anomalous vertical velocity are displayed in Fig. 4. The static stability is defined by  $\frac{g}{\theta} \frac{\partial \theta}{\partial z}$ , where  $g$  is the acceleration

due to gravity ( $9.81 \text{ ms}^{-2}$ ) and  $\theta$  is potential temperature [K]. Variables are averaged over the tropical band between  $20^\circ \text{ S}$  and Equator, where the convective activity and the upwelling in the TTL showed clear changes during the SSW event (Fig. 3).

Temperatures started to decrease in the LS and the upper TTL (above 17.5 km) around 15 January, but increased in the troposphere. The region of weak negative temperature trend then propagated downward in the TTL during period (ii) and (iii). The zone of weaker static stability also descended through the TTL following the negative temperature trend. The anomalous vertical velocity (Fig. 4b) showed a strengthening of the upwelling propagating from the LS to the UT in period (ii), similar to the temperature tendency, but with enhancement in the deep troposphere during period (iii), when total column integrated cloud fraction was the largest (Fig. 2c).

To understand the different characteristics of the variations in temperature tendency and vertical velocity, we investigated the major terms (the term on the LHS and first two terms on the RHS of the following equation) of the thermodynamic equation in the Transformed Eulerian Mean (TEM) framework (Andrews et al., 1987);

$$\frac{\overline{\partial T}}{\partial t} = -N^2 \overline{w^*} \left( \frac{\overline{P_0}}{P} \right)^{-\kappa} + \overline{\text{DH}} - \frac{\overline{v^*} \overline{\partial T}}{a \partial \phi} - \frac{1}{\rho_0} \frac{\partial}{\partial z} \left[ \rho_0 \left( \frac{\overline{v'T'}}{a \frac{\partial T}{\partial z}} \frac{\partial T}{\partial \phi} + \overline{w^* T'} \right) \right].$$

Here  $T$  is temperature,  $N$  is Brunt–Väisälä frequency,  $w^*$  ( $v^*$ ) is the vertical (meridional) component of residual velocity, DH is diabatic heating rate (cloud microphysics and radiation),  $P(P_0)$  is (reference) pressure and  $\kappa$  is  $R/C_p$  (0.286), (where  $R$  is the gas constant and  $C_p$  is the specific heat at constant pressure). The third and fourth terms of the RHS of the equation are meridional advection and diffusion by waves and eddies, respectively.

The temperature tendency agrees quite well with the adiabatic cooling rate at 20 km (LS) (Fig. 4d). The evolution of these curves matches that of the eddy heat flux ( $\overline{v'T'}$ ) at 100 hPa averaged over the extratropical NH (Fig. 4c). Note also that the radiative diabatic heating rate remains almost constant at  $+0.5 \sim 0.6 \text{ K day}^{-1}$ , which may balance the third and fourth terms on the RHS of the above thermodynamic equation.

At the bottom part of the TTL (14.3 km in Fig. 4e), the adiabatic and diabatic heating were almost in balance, while the adiabatic cooling associated with the stratospheric upwelling drove

the temperature tendency at greater heights (above 14.3 km). Meanwhile, the heat balance in the TTL was also related to the longitudinal distributions of diabatic heating due to cloud microphysics and solar (shortwave) radiation.

Figure 5a and b shows the horizontal variation of the latent heating rate due to cloud microphysics as an indicator of cloud formation at the bottom of the TTL (14.3 km). Figure 5c and d shows the longitude–height section of vertical velocity averaged over 20° S–Equator. Left and right panels display the averages during periods (i) and (ii), respectively. In period (i), before the SSW event started, clouds formed mainly along the intertropical convergence zone (ITCZ) and over the southwestern Pacific region in the UT, which extended up to the TTL. In the TTL, clouds were also found over equatorial Africa and South America.

When the LS in the southern tropics started to cool in period (ii), convective activity in the TTL shifted southward and increased along the latitudes 20–10° S (Fig. 3b), particularly over the southwestern Pacific, southwestern Indian Ocean, and coastal regions of South America. The development of convective intensity and depth from the period (i) to (ii) is apparent in the pressure–longitude section of vertical velocity, especially over the western Pacific and the western Indian Ocean (Fig. 5c and d). The upward motion above the TTL, especially at 85 hPa, became dominant in period (ii).

Thus, diabatic heating and vertical motion associated with deep convection with zonally asymmetric distributions influenced the heat balance in the UT and TTL, whereas in the LS the effects of zonally symmetric vertical motion were dominant (as indicated by Figs. 4d, e and 5c, d).

## 4 Summary and discussion

The present study investigates stratospheric dynamical impacts on the tropical tropospheric convection during a SSW event from the view point of thermo-dynamic balance in the TTL using simulation data from a global non-hydrostatic model (NICAM). NICAM can provide the physically consistent data for vertical velocity and diabatic heating, which are difficult to observe. The model reproduced the observed processes during SSW: convective activity in the

tropical SH was enhanced following an amplification of extratropical planetary wave activity in the winter NH.

The stratosphere–troposphere connection was detected in the correlation coefficient between the tropical stratospheric upwelling and the vertical velocity in the tropical troposphere. The highest correlation was found in the SH 20–10° S latitudinal band (Fig. 3a), where the upwelling branch of BD circulation in the summer hemisphere is located for this case. Particularly large variations were found over the southwestern Indian Ocean, southwestern Pacific, and coastal regions (Fig. 5).

Although upwelling in the troposphere occurs following that of the stratosphere, they were not produced by the same process (Fig. 4). Vertical velocity over the wide tropics (e.g. 30° S–30° N) in the LS is mainly driven by extratropical planetary waves (Ueyama et al., 2013; Abalos et al., 2013), but that in the UT was affected mainly by deep convection in a more restricted zone around 20° S–10° S.

The enhanced upwelling in the LS associated with the SSW event can intrude deeper into the TTL and cause adiabatic cooling of the upper TTL. Accordingly, the unstable situation in the TTL causes the convection higher, that means the existing convections were enhanced mainly over the western Pacific the western Indian Ocean and the south western Pacific, shown in Fig. 5. The paper by Kodera et al. (personal communication given at the AGU fall meeting, 2013) found the overshooting clouds increase and higher after the SSW events in 2009 and 2010 by using the observational data.

Following an increase of stratospheric meridional circulation, the cirrus clouds in the TTL and convection in the southern hemisphere were enhanced after the onset of SSW. It can be shown that the enhanced adiabatic cooling in the LS destabilizes the underlying TTL, promoting ice cloud formation over the TTL (Fig. 3b and c) and stronger convective activity. Li and Thompson (2013) showed that month-to-month variation of the clouds around the tropopause was attributable to changes in the static stability caused by cooling in the tropical LS associated with the BD circulation.

However, the convective clouds around 20–10° S latitude in period (ii) extended over the whole troposphere, and appeared as longitudinally localized (mesoscale) convective activities.

In particular, the strong upwelling near Madagascar and south western Pacific in Fig. 5b and d was associated with a tropical cyclone. These results suggest that more detailed analysis in time and space is needed to understand the interaction between the global scale stratospheric circulation and the mesoscale organization of cloud in the troposphere.

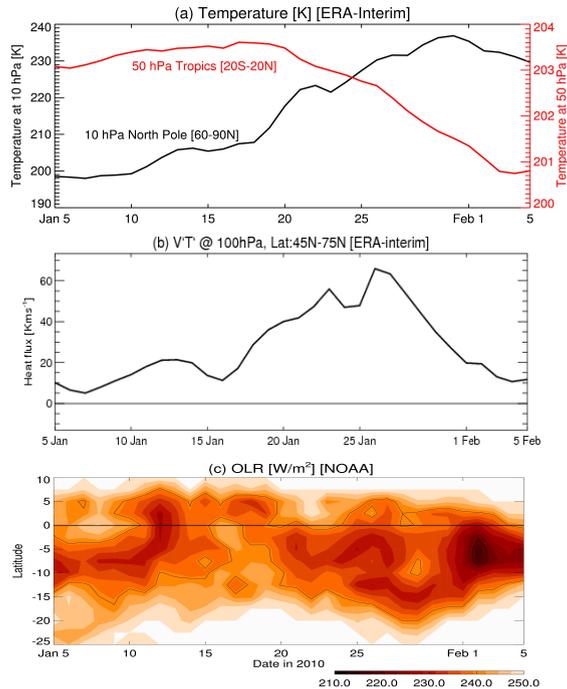
- 5 *Acknowledgements.* The NICAM simulation was performed on the Earth Simulator. The authors would like to thank Chihiro Kodama at the Japan Agency for Marine-Earth Science and Technology (JAM-STE) for his useful comments. This study was supported by JSPS KAKENHI Grant-in-Aid for Young Scientists (B), number 23710025 and JSPS KAKENHI Grant-in-Aid for Scientific Research (C), number 25340010.

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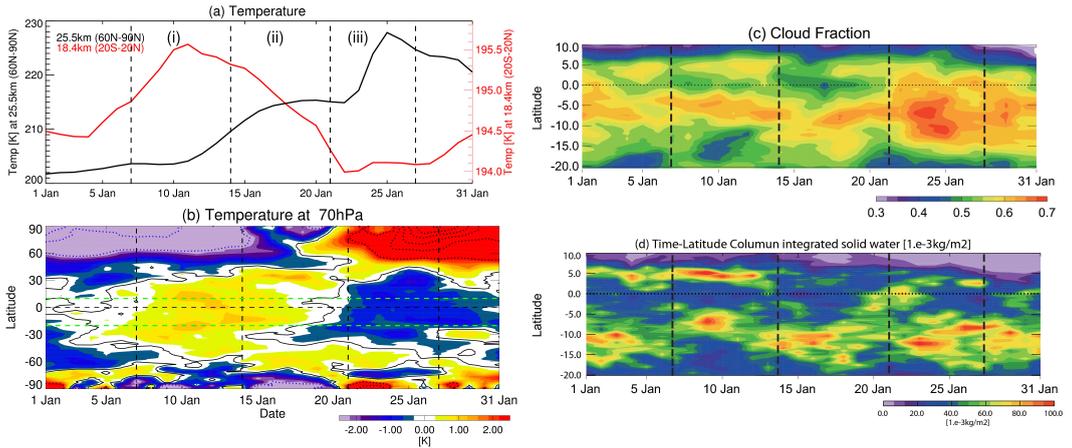
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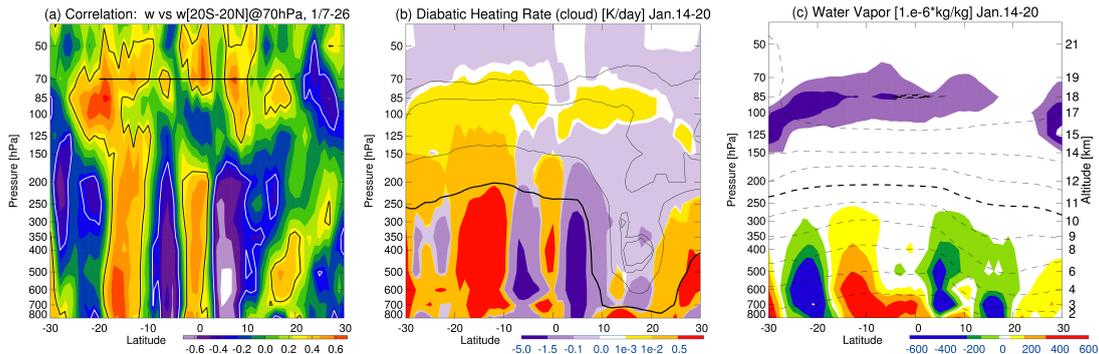
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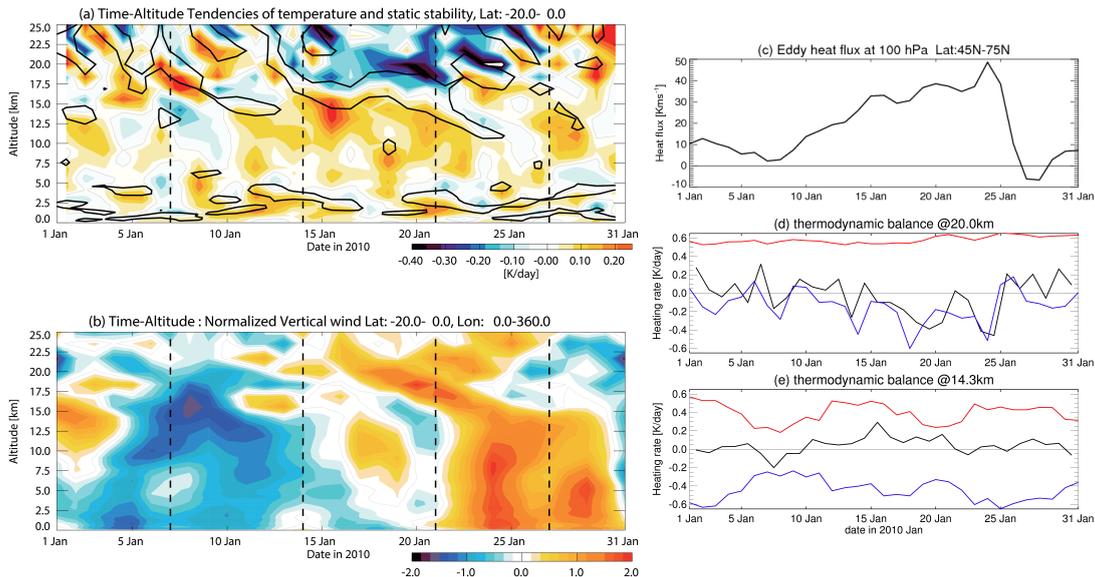
**Fig. 1.** (a) Time series of observed zonal mean temperature [K] in the northern high latitude at 10 hPa (black line) and tropical at 50 hPa (red line) from 5 January to 5 February, 2010. Note the different temperature scales for the polar (left) and tropical (right) regions. (b) Time series of observed zonal mean eddy heat flux [ $\text{K ms}^{-1}$ ] at 100 hPa averaged between  $45^\circ \text{N}$  and  $75^\circ \text{N}$ . (c) Time–latitude ( $25^\circ \text{S}$ – $10^\circ \text{N}$ ) section of NOAA observed zonal mean OLR [ $\text{W m}^{-2}$ ]. The horizontal line indicates the equator.



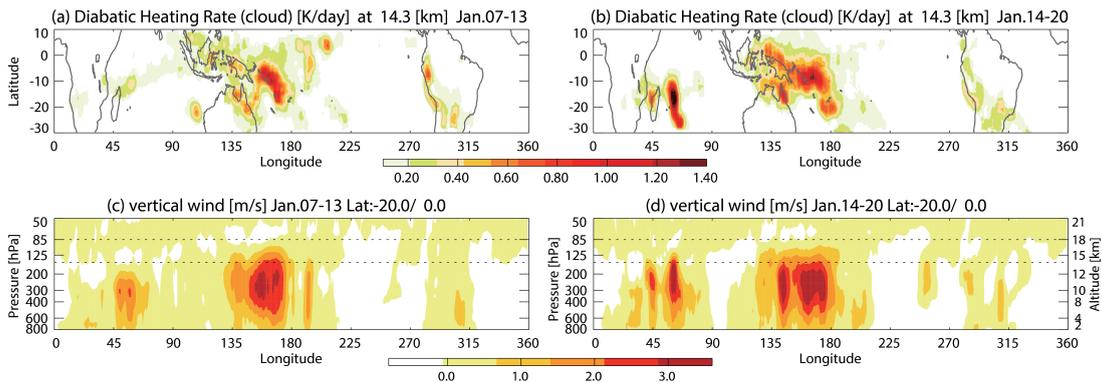
**Fig. 2.** (a) Same as Fig. 1a but for simulated zonal mean temperature from 1 to 31 January 2010. (b) Time–latitude ( $90^{\circ}\text{N}$ – $90^{\circ}\text{S}$ ) section of simulated temperature anomaly [K] at 70 hPa with respect to the January 2010 monthly mean at each latitude. The solid contour indicates zero anomaly and the dotted contours indicate  $-2.5$  K and  $2.5$  K. The light green horizontal dashed lines indicate the latitudinal region of (c) and (d). (c) Time–latitude ( $20^{\circ}\text{S}$ – $10^{\circ}\text{N}$ ) section of simulated zonal mean column integrated cloud fraction. (d) Same as (c) but for column integrated ice cloud [ $10^{-3} \text{ kg m}^{-2}$ ]. The four vertical dashed lines in (a–d) are 7, 14, 21 and 27 January; the periods between the lines are the (i) initial, (ii) transition and (iii) mature phases, respectively. The horizontal dashed lines in (b, c and d) indicate the equator.



**Fig. 3.** (a) Latitude–pressure sections of correlation between the vertical wind averaged over  $20^{\circ}$  S– $20^{\circ}$  N at 70 hPa (thick line) and vertical wind at each level and latitude for the period of 7–26 January. Latitude–pressure sections of (b) latent heating rate due to cloud formation [ $\text{K day}^{-1}$ ] (contours) and (c) the water vapor concentration [ $10^{-6} \text{ kg kg}^{-1}$ ] (dashed contours) averaged between 14 and 20 January. The thin contours in (b) are values of  $10^{-3} \times 10^i$  ( $i = 0, 1, 2$ ) [ $\text{K day}^{-1}$ ]. The thick contour is 1.0 [ $\text{K day}^{-1}$ ]. The dashed contours in (c) are values of  $10^4 \times 10^{i/3}$  ( $i = 0, 1, 2, \dots, 10$ ) [ $10^{-6} \text{ kg kg}^{-1}$ ]; the thick dashed contours indicate 4.6 and 100 [ $10^{-6} \text{ kg kg}^{-1}$ ]. Anomalies with respect to the 7–13 January average are shown by color shading. The light and heavy purple in (c) indicate values of  $-0.3$  and below,  $-0.6$  and below.



**Fig. 4.** Time–altitude section of **(a)** temperature tendency [ $\text{K day}^{-1}$ ] (color shading) and tendency of static stability with value of  $-0.02 [10^{-4} \text{ s}^{-2} \text{ day}^{-1}]$  shown by black contours, and **(b)** anomalous normalized vertical velocity averaged between  $20^\circ \text{ S}$  and Equator from 1 to 31 January 2010. The normalized values are obtained by dividing by the standard deviation at each altitude. **(c)** Time series of zonal mean eddy heat flux [ $\text{K m s}^{-1}$ ] at 100 hPa averaged between  $45^\circ \text{ N}$  and  $75^\circ \text{ N}$ . **(d, e)** show the time series of thermodynamic balance at 14.3 and 20.0 km averaged between  $20^\circ \text{ S}$  and Equator. The black, red and blue lines indicate temperature tendency ( $\partial T/\partial t$ ), diabatic heating rate due to cloud microphysics and radiation, and adiabatic heating rate due to residual vertical velocity ( $-N^2\omega^*$ ), respectively.



**Fig. 5.** Maps of diabatic heating rate from cloud microphysics [ $\text{K day}^{-1}$ ] at 14.3 km (**a**, **b**) averaged over 7–13 January (left, **a**) and 14–20 January (right, **b**). The panels (**c**, **d**) show longitude–pressure sections of vertical wind [ $\text{m s}^{-1}$ ] averaged between  $20^\circ$  S and Equator on 7–13 and 14–20 January, respectively. The vertical wind data are smoothed by applying a weighting function of five-degree-longitude width. The horizontal dotted lines in (**c**, **d**) show the 85 hPa and 150 hPa pressure levels for reference.