Intraseasonal Variability of the Zonal-Mean Tropical Tropopause Height

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ABSTRACT

Intraseasonal variability of the zonal-mean tropical tropopause height is shown to be modulated by localized tropical convection. Most of this convective activity is identified as being part of the Madden–Julian oscillation. While the convection is highly localized over the Pacific warm pool, a large-scale circulation response to the convective heating rapidly warms most of the tropical troposphere and cools most of the lowest few kilometers of the tropical stratosphere. These changes in temperature fields raise the tropical tropopause at most longitudes within 10 days of the convective heating maximum.

1. Introduction

While theoretical models of radiative–convective equilibrium (Manabe and Strickler 1964; Held 1982; Thuburn and Craig 2000) have been considered as successful at explaining the formation of the tropical tropopause, analyses of local radiosonde data have revealed that the temperature structure does not entirely conform to the radiative–convective equilibrium model. In a typical sounding, the temperature profile is consistent with moist convective adjustment only up to about 11–13 km, well below the tropopause level of 15–16 km (Highwood and Hoskins 1998; Gettelman and de Forster 2002). This suggests that tropical tropopause height is likely being modulated by factors other than the convection itself, such as large-scale upwelling in the tropical stratosphere (Yulaeva et al. 1994).

Numerical model simulations by Boville (1984), Thuburn and Craig (2000), and Salby and Callaghan (2004) showed that adiabatic cooling associated with mean upwelling in the lower stratosphere could raise the tropical tropopause by up to 1 km. This tropical upwelling has been shown to be important both for the seasonal cycle (Randel et al. 2000; Seidel et al. 2001; Gettelman and de Forster 2002) and for interannual variability (Randel et al. 2000; Wong and Wang 2003) of the zonal-mean tropical tropopause height. Since the upwelling is driven primarily by wave breaking in the extratropical stratosphere (Rosenlof 1995; Randel et al. 2002), the resulting tropical tropopause height change is a nonlocal response to the midlatitude eddy driving.

Compared with the annual and interannual timescale variability, intraseasonal variability of the tropical tropopause height \( Z_T \) has started to receive attention only recently. It is shown, for intraseasonal time scales, that the zonal distribution of \( Z_T \) is highly affected by local processes at the troposphere, such as convection and convectively driven waves (Zhou and Holton 2002; Randel et al. 2003). In particular, Zhou and Holton (2002) showed that in response to Madden–Julian oscillation (MJO; Madden and Julian 1994) convection the cold point tropopause rapidly cools to the east of the convection. They attributed this cooling to the presence of a Kelvin wave. On the other hand, Thompson and Lorenz (2004) indicate that the daily evolution of the Northern Hemisphere (NH) annular mode significantly modifies tropical lower stratospheric temperature (the proximity of these lower stratospheric temperature changes to the tropopause will determine whether this process impacts \( Z_T \)). Since zonal-mean features are emphasized in Thompson and Lorenz (2004), taken together with the previous studies, it is unclear whether the intraseasonal variability of the zonal-mean \( Z_T \) ([\( Z_T \)] hereafter) is controlled primarily by local or nonlocal processes.

The primary goal of this study is to examine the relative importance of local versus nonlocal forcing of the intraseasonal variability of \( Z_T \). The data and methodology used in this study are described in section 2. Sec...
tion 3 examines a possible link between tropical and extratropical tropopause height on intraseasonal time scales. The physical processes of $[Z_T]$ fluctuation are described in section 4. Finally, conclusions follow in section 5.

2. Data and approach

To examine $[Z_T]$, we use the global National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset (Randel et al. 2000; Kiladis et al. 2001; Wong and Wang 2003; Santer et al. 2003). Although it would have been preferable to have used local radiosonde data, as in most tropopause studies, we chose not to use these data because of its sparseness and also because of possible inconsistencies amongst the data. Although the global reanalysis dataset is known to suffer a negative bias in $Z_T$ (Randel et al. 2000), it is still adequate for investigating the temporal variability of $Z_T$ (Randel et al. 2000; Kiladis et al. 2001).

There are two competing definitions for the tropopause, that is, the cold point and the lapse-rate tropopause. The cold point tropopause is defined as the level of lowest temperature. This definition is appealing because chemical constituents transported into the stratosphere depend on the coldest temperature along their path (Holton et al. 1995; Highwood and Hoskins 1998). However, the vertical resolution of the NCEP–NCAR reanalysis dataset is too coarse to accurately resolve the cold point tropopause. Thus, we adopt the conventional definition that the height of the tropopause is the lowest level at which the temperature lapse rate remains below 2 K km$^{-1}$ for an upward distance of 2 km [World Meteorological Organization (WMO) 1957].

The daily NCEP–NCAR reanalysis dataset covered by this study spans the time period from 1 January 1979 to 31 December 2004. All reanalysis fields have been interpolated from sigma to height coordinates, with a 0.5-km interval, by using a cubic spline. The zonal-mean tropopause height $[Z]$ is then identified in accordance with the above WMO definition. Although not shown, the resulting time series of $[Z]$, averaged from 15°S to 15°N, is very similar to that of the zonal-mean tropical tropopause pressure provided by NCEP. The correlation coefficient between the two time series is $-0.95$. The intensity and location of convection is estimated from the daily outgoing longwave radiation (OLR), archived by the National Oceanic and Atmospheric Administration (NOAA). For this study, the seasonal cycle, which is defined as the long-term calendar day mean, is removed for all variables.

All analyses in this study are based on linear regression against the zonal-mean tropical tropopause height $([Z_T])$, where the angle bracket denotes latitudinal average from 15°S to 15°N:

$$x'(\tau) = r(x'(\tau), [Z_T])'\frac{\sigma_{x'}}{\sigma([Z_T])'}([Z_T])',$$

where the prime denotes a deviation from the seasonal mean for each year. By subtracting the seasonal mean for each year, we also remove the interannual variability. The seasons are divided into 3-month intervals. For example, the NH winter/SH summer is defined as December, January, and February (DJF hereafter). In (1), $x$ is the variable of interest, $r(A(\tau), B)$ is the linear correlation coefficient between $A$ and $B$ at the $\tau$-day lag, and $\sigma_x$ is the standard deviation of $x$. The form of $x'(\tau)$ associated with a one standard deviation value in $([Z_T])'$ is then

$$x'_\tau(r) = r(x'(\tau), [Z_T])'\sigma_{x'},$$

$$= \frac{1}{\sigma([Z_T])'} \sum_{\tau=1}^{N_d} (x'(\tau)[Z_T])'(t) - \bar{x}'_{[Z_T]}$$

where $N_s$ and $N_d$ are, respectively, the total number of years and the number of days in a given season. For conciseness, the regressed variable $x'_\tau$ will be referred to as the anomaly and denoted only with prime. Note that both the seasonal cycle and the interannual variability are removed from all variables.

Statistical significance is determined from the correlation field. Following Oort and Yienger (1996), the number of degrees of freedom (NDOF) is estimated as

$$\text{NDOF} = N \left[ 1 + \frac{2}{N_s} \sum_{\tau=1}^{N_s} (N - \tau)r_{x'(\tau)[Z_T]}' \right]^{-1},$$

where $N_s$ is the sample size equal to $N_s \times N_d$. The quantity $r_{x'(\tau)}$ is an autocorrelation function at lag $\tau$. For most fields, the NDOF is about 100 to 150 in the Tropics and increases by more than twofold in the extratropics. The increase in the NDOF results from the fact that the decorrelation time scale of the variable $x'$ is generally shorter in the extratropics than in the Tropics.

3. Relationship between tropical and extratropical $[Z]$}

On the seasonal and interannual time scales, there is an out-of-phase relationship in $[Z]$ between low and high latitudes (Yulaeva et al. 1994; Randel et al. 2000). Therefore, we first examine the extent to which the
intraseasonal variability of $[Z_T]$ is associated with that of extratropical $[Z]$. Figure 1 shows the time evolution of $[Z]$ for each season. A statistically significant relationship between $([Z_T])$ and the subpolar $[Z]$, at the 95% confidence level, is found during the SH spring (Fig. 1b) and the NH winter (Fig. 1c). This seesaw-like variability is qualitatively similar to that observed for the interannual variability, suggesting that the intraseasonal variability of $[Z_T]$ may be affected by the nonlocal forcing. Upon examining the NH winter more closely, it can be seen that the seesaw-like variability is indeed associated with stratospheric temperature anomalies (Fig. 2a).

Although the nonlocal forcing seems to play some role for the NH winter and the SH spring, Fig. 1 also indicates that the $[Z_T]$ anomalies themselves show no appreciable seasonal differences. For all seasons, fluctuations of $[Z_T]$ take place over a broad latitudinal band with sharp boundaries at around 30° in both hemispheres, which correspond to the poleward end of the two Hadley cells. In addition, regardless of the stratospheric signal, $\sigma_{([Z_T])}$ is about 100 m for all seasons. These results suggest that, for the intraseasonal time scale, the $[Z_T]$ anomalies have a strong link to local forcing in the tropical troposphere. For instance, during the boreal summer [June–July–August (JJA)], the temperature anomalies associated with $([Z_T])$ occur only in the Tropics (Fig. 2b).

It is noteworthy that $([Z_T])$ is significantly correlated with the temperature in the lower troposphere (see the shading at the surface in Fig. 2). The maximum correlation coefficient between $([Z_T])$ and $[T]$ at 1 km is 0.45 for DJF and 0.60 for JJA. As an example, the time series for $([Z_T])$ and the 1-km $[T]$ are presented in Fig. 3 for the 1989/90 winter. For this particular time period, the correlation between the two time series is greater than 0.8. These results provide additional evidence of tropospheric control of the intraseasonal variability of $([Z_T])$. If $([Z_T])$ is controlled by remote forcing in the extratropical stratosphere, it is unlikely for $([Z_T])$ to be associated with near-surface temperature fluctuations. Furthermore, the positive correlations between $([Z_T])$ and near-surface temperature are consistent with radiative–convective equilibrium theory (e.g., Manabe and Strickler 1964).

The above results suggest that for all seasons it is processes local to the Tropics that directly affect the intraseasonal variability in $[Z_T]$, although nonlocal stratospheric forcing also influences the tropical tropopause height. The latter possibility may be associated with the stratospheric annular mode, as shown by Thompson and Lorenz (2004). While the link between the nonlocal stratospheric forcing and tropical processes warrants further investigation, in this study, we focus on tropical processes that are shown to be robust throughout all seasons.
4. Tropical processes

Building upon the above results, this section investigates the physical processes associated with the local tropical forcing that drives $[Z_T]$ fluctuations. Because the elevation of $[Z_T]$ is associated with large-scale warming in the upper troposphere (UT) and cooling in the lower stratosphere (LS) (Fig. 2), we first examine the horizontal structure of the temperature field in these regions. For this purpose, we define the UT warming and LS cooling as corresponding to the levels of the maximum and minimum temperature anomalies in Fig. 2, which occur at approximately 14 and 18 km, respectively. Including these two temperature fields, all variables displayed are slightly smoothed by truncating them at zonal wavenumber 5. Although not shown, the results described below were found to be insensitive to the choice of smoothing operator.

a. UT fields during DJF

The daily evolution of the UT temperature anomalies, shown in the left column of Fig. 4, indicate that the UT warming undergoes large structural changes during the 15-day period preceding the maximum $[Z_T]$. As seen in Fig. 4d, the UT warming at the lag 0 day is largely homogeneous in the zonal direction with a short break at 120°W. This overall warming, however, is initiated within a small region over the western Pacific (Fig. 4a). This anomaly, denoted by A, grows as it slowly moves eastward (Figs. 4b,c). At the same time, two other well-defined anomaly centers, denoted by B and C, start to appear to the east and west of A (Figs. 4b,c). Another anomaly D is also observed at lag 0 day. These anomalies quickly spread out and eventually join A, resulting in a warming throughout the entire Tropics (Fig. 4d).

The above results, warming over A followed by B, C, and D, suggest that the large-scale circulation excited by western Pacific convection may drive the temporal evolution of the UT temperature anomalies. The anomalous OLR and vertical velocity fields during the same analysis period (Fig. 5) support this possibility. The OLR anomaly indicates that convective activity is enhanced over anomaly A but is suppressed over anomalies B, C, and D (Figs. 5a–d). Consistent with the
OLR fields, the vertical velocity fields show anomalous rising over A, but with sinking over the other anomalies (Figs. 5e–h).

The results obtained thus far suggest that the large-scale circulation, responding to localized convection, rapidly warms the entire Tropics. In particular, while A arises from diabatic warming associated with the convective activity, the other three anomaly centers develop as a result of dynamic warming caused by the large-scale circulation. To evaluate this process, we consider the thermodynamic energy equation:

\[
\frac{\partial T}{\partial t} = Q_{\text{DYN}} + Q_{\text{DIA}}, \tag{3}
\]

where

\[
Q_{\text{DYN}} = -\nabla \cdot (TV) + TV \cdot \nabla - w \left( \frac{\partial T}{\partial z} + \frac{g}{C_p} \right)
\]

\[+ K_n \nabla^4 T.\]

The notations are standard. In (3), \(Q_{\text{DYN}}\) and \(Q_{\text{DIA}}\) denote the dynamic and the diabatic heating, respectively. Again, all terms in (3) are averaged from 15°S to 15°N and then regressed against \(\langle Z \rangle\) as in (2). Thus, the result shown in Figs. 6 and 7 are all for anomalies. Because the residual of the budget is taken as the \(Q_{\text{DIA}}\) term, we caution that this term may include numerical errors. The errors, however, seem to be reasonably
As can be seen in Figs. 6c and 6d, for example, the derived $Q_{\text{DIA}}$ closely resembles the OLR field; not only the spatial pattern but also the relative amplitude of the anomalies share similarities. Because the OLR field is independent of the budget calculation, it is highly unlikely that the derived $Q_{\text{DIA}}$ field grossly misrepresents the actual diabatic heating anomalies.

The budget analysis indicates that the warming starts over the western Pacific (Fig. 6a), consistent with the temperature anomaly field shown in Fig. 4a. This initial warming, denoted by A, is caused by the combined effect of anomalous diabatic warming ($Q_{\text{DIA}}$) and dynamic warming ($Q_{\text{DYN}}$) (Figs. 6b,c). A close examination of Figs. 6b and 6c indicates that ($Q_{\text{DIA}}$) is stronger between 120°E and 180°, whereas ($Q_{\text{DYN}}$), while still weak, plays the major role to the east of the date line between 180° and 140°W. A comparison of ($Q_{\text{DIA}}$) (Fig. 6c) with the OLR anomaly (Fig. 6d) further reveals that the anomalous diabatic heating is closely associated with convective activity. Although not shown, the dynamic warming east of the date line is driven by horizontal temperature advection. As such, the anomaly center A is initiated by convective heating, but it expands farther eastward by horizontal heat transport. This result is consistent with Figs. 5a to 5d, which

\[1\] Recalling that the budget is for anomalies, the negative values in ($Q_{\text{DIA}}$) do not necessarily indicate diabatic cooling but also represent anomalously weak diabatic heating, which reflects reduced convective activities.
show that the minimum OLR anomalies occur slightly west of the anomaly center A. In Fig. 6a, three other temperature anomaly centers are also observed between lag $-10$ and lag $-5$ days. These anomalies are all driven by dynamic warming (Fig. 6b). Although not shown, this dynamic warming is mainly due to large-scale descent with a weak but nonnegligible contribution from horizontal temperature advection.

The behavior of the OLR anomalies associated with $[Z_T]'$ is reminiscent of the MJO. Figure 6d shows that the OLR anomalies in the eastern hemisphere propagate at a speed of $5-6$ m s$^{-1}$. The recurrence interval for this eastward-propagating OLR anomaly is about $40-50$ days. These characteristics are consistent with the MJO, suggesting that convective activity associated with the

Fig. 6. (a) The time tendency of tropical mean UT temperature anomalies during DJF, (b) the contribution from dynamic heating, (c) the contribution from diabatic heating, and (d) the tropical mean OLR anomalies. (e) Same as in (a) but for the LS temperature anomalies. The contour intervals are $0.01$ K in (a)–(c), $1$ W m$^{-2}$ in (d), and $0.02$ K in (e). In all figures, zero lines are omitted and statistically significant values at the 99% (95%) confidence level are shaded in dark (light) gray. Note that the longitudinal locations of A to D are exactly the same as those in Figs. 3 and 4.

Fig. 7. The same as in Figs. 6a–d but for JJA.
intraseasonal $[Z_T]$ variability is part of the MJO. This result indicates that MJO convection not only modulates the local $Z_T$ through convective heating (Kiladis et al. 2001; Zhou and Holton 2002) but also the zonal-mean $Z_T$ through convectively induced large-scale circulations (e.g., Kiladis et al. 2005).

b. LS fields during DJF

The LS temperature anomaly is displayed in the right column of Fig. 4. Its overall pattern closely resembles that of the UT temperature anomalies (left column) but with opposite sign. As for the UT temperature anomalies, there are four local extrema in the LS temperature anomaly fields. More importantly, these four anomaly centers are collocated with the underlying UT anomaly centers (see also Fig. 6e). Since much of the UT temperature anomalies are shown to be associated with both a local convection and large-scale circulation driven by this convection, it is highly probable that the LS temperature anomalies also arise from the same convection and large-scale circulation. This is consistent with previous studies, which show that LS cooling can occur not only directly above the convection (Johnson and Kriete 1982; Reid 1994; Sherwood et al. 2003; Kiladis et al. 2005) but also distances away from the convection (e.g., Highwood and Hoskins 1998; Zhou and Holton 2002). The latter far-field response may involve either Gill (1980)-type tropospheric wave response (Highwood and Hoskins 1998) or vertically propagating equatorial Kelvin waves (Tsuda et al. 1994; Randel and Wu 2005).

c. UT/LS fields during JJA

A parallel analysis is performed for the boreal summer, and the results are summarized in Fig. 7. Since maximum UT temperature and OLR anomalies during JJA occur in the Northern Hemisphere subtropics (e.g., Fig. 8e), the analysis fields are averaged from 10°S to 20°N rather than from 15°S to 15°N. As for the boreal winter, the tropospheric warming is initialized in the western Pacific and spreads over the entire Tropics within 10 to 15 days (Fig. 7a). The initial warming in the western Pacific is once again caused by convectively driven diabatic heating (Figs. 7c,d). This anomaly is further maintained by horizontal temperature advection and adiabatic heating associated with weakened convective activity (Fig. 7b). Again, similar to the DJF result, the warming to the east and west of the convection is due to dynamic heating (Fig. 7b), driven by the large-scale circulation. As such, the overall physical picture is qualitatively similar to that seen in the boreal winter.

One difference between the two seasons is the location of the maximum UT warming relative to the center of the convection. During DJF, the initial warming occurs to the east of the convection (cf. Figs. 6a and 6d), whereas during JJA, the initial warming occurs to the west of the convection (cf. Figs. 7a and 7d). This seasonal difference results from the dynamic heating. Figure 8 shows the temporal evolution of the anomalous streamfunction ($\psi'$) and wind fields in the UT from lag $-15$ to $-5$ days. Superimposed thick contour lines denote the location and intensity of the convective activity. It is seen that zonal flow anomalies during DJF are to a large degree westerly (Figs. 8b,c). The easterly anomalies west of the convection are almost negligible. In contrast, during JJA, the easterly anomalies to the west of the convection are quite strong (Figs. 8e,f). They are caused by the divergent flow associated with the Rossby gyres. Although strong westerly anomalies are also observed to the east of the convection, the associated heat transport from the convective region is rather minor.

Taking the above results together, it is concluded that the heating from the convection over the western Pacific is advected primarily eastward (westward) during DJF (JJA) through the influence of Kelvin (Rossby) waves. However, it should be emphasized that, in both seasons, the warming at a greater distance from the convection is associated with the eastward propagating Kelvin waves (e.g., cf. Figs. 4a–c, 6a, and 8a–c for DJF and Figs. 7a and 8d–f for JJA). For instance, the warming at A in Fig. 4 is followed at a later time by warming at B, C, and D. This is consistent with Kelvin wave propagation (Figs. 8a–c), suggesting that the rapid warming over the entire Tropics is due primarily to the convectively driven Kelvin waves and associated adiabatic warming. Reflecting the fact that the Kelvin wave plays the dominant role for the modulation of $\langle[Z_T]\rangle'$, Fig. 8 shows 99% (and 95%) significant level for the streamfunction coherence, only for the Kelvin wave.

d. Power spectra

The characteristics of the intraseasonal variability of $\langle[Z_T]\rangle'$ are further examined with power spectral analysis. Figures 9a and 9d show the intraseasonal power spectra of $\langle[Z_T]\rangle'$ for DJF and JJA, respectively. These power spectra are obtained by calculating the average of individual power spectra for each of 25 DJF and 26
Fig. 8. The anomalous UT streamfunction $\psi'$ and wind vectors during (a)–(c) DJF and (d)–(f) JJA from lag –15 to lag –5 days. The contour interval is $0.5 \times 10^6$ m$^2$ s$^{-1}$, and zero lines are omitted. Statistically significant values at the 99% (95%) confidence level are shaded in dark (light) gray. Negative OLR anomaly values less than $-3.5$ W m$^{-2}$ are superimposed in the figures with thick solid lines. The wind vectors, which are statistically significant at the 99% (95%) confidence level, are plotted in black (dark gray).
JJA seasons. For each season, the linear trend is removed and tapering is applied. In addition to the power spectra, Fig. 9 also displays the red noise spectra and the corresponding 95% confidence level. Assuming 2 degrees of freedom for each season, the confidence limits are determined by setting the number of degrees of freedom to 50 for DJF and 52 for JJA.

For JJA, the $<[Z_T]>'$ power spectra exhibit a statistically significant spectral peak at around 45 days (Fig. 9d). Although lacking a distinct peak, DJF $<[Z_T]>'$ spectra also show a significant power at 45 days (Fig. 9a). This period overlaps with the period of statistically significant peak in the power spectra of OLR anomalies over the Pacific warm pool (Figs. 9b,e). By referring to the location of maximum convection in Fig. 8, the warm pool OLR anomaly (OLRwp) is defined as an average from 120°E to 180° and from 15°S (10°S) to 15°N (20°N) for DJF (JJA). The broad band of statistically significant peaks in the (OLRwp)' power spectra matches those of the MJO. Not surprisingly, the lag correlation between $<[Z_T]>'$ and (OLRwp)' for the raw data (black) and 30-day low-pass-filtered data (gray) in each season. Here, the negative time lag corresponds to (OLRwp)' leading $<[Z_T]>'$. The two horizontal lines in (c) and (f) denote the values that are statistically significant at the 99% confidence level.

The correlation coefficients increase rather substantially if the calculations are performed with a 30-day low-pass filter applied to both time series (see the gray curves in Figs. 9c,f).
of (\(\text{OLR}^{\text{wp}}\))' is qualitatively similar to that of \(\langle Z_T \rangle'\) but with a 5- to 15-day time lag.

5. Conclusions

For intraseasonal time scales, zonal-mean tropical tropopause height variability is driven primarily by local convective heating and the large-scale circulation response to this convective heating. Adiabatic processes associated with the large-scale circulation rapidly heat the entire tropical troposphere and cool the lower stratosphere. Because the tropopause is raised by UT warming/LS cooling, at least for intraseasonal variability, it is the large-scale response to the convective heating, rather than the local convection itself, that appears to be central in modulating the zonal-mean tropical tropopause height.

This type of large-scale adiabatic UT warming and LS cooling was suggested by Highwood and Hoskins (1998) in the context of the Matsuno (1966) and Gill (1980) models. Apparently, however, their picture explains only part of the observed anomalous temperature structure. In the Tropics, the initial \(\psi'\) and wind vector fields (Fig. 8a) are qualitatively similar to those for the Gill-type response to localized heating (e.g., Jin and Hoskins 1995), with a pair of Rossby waves to the west and a Kelvin wave to the east of the localized heating. Comparing the \(\psi'\) fields in Fig. 8 with the corresponding temperature anomalies in Fig. 4, the latitudinal broadening of center A, seen throughout Figs. 4a to 4d (also in Figs. 4e–h), is associated with the Rossby wave pair. This behavior conforms to the picture summarized in Fig. 9 of Highwood and Hoskins (1998). On the other hand, centers B, C, and D first emerge at the leading edge of the Kelvin wave front, rather than directly underneath the Kelvin wave westerly maximum (again see Fig. 9 of Highwood and Hoskins). Since the leading edge of the Kelvin wave also seems to be associated with a poleward-propagating Rossby wave train (this is particularly clear in Fig. 8a), it is possible that other less well-known processes, such as a Rossby wave-induced mean meridional circulation (Boehm and Lee 2003), may also play an important role for generating the UT warming/LS cooling centers and thus for the zonalization of the UT warming/LS cooling.

Regardless of the precise mechanism, because the Rossby radius of deformation is very large in the Tropics, it may not be surprising that the response to a localized heating is felt by the entire Tropics. On the other hand, one cannot rule out that the global-scale response may be due to the construction of our data analysis. Because the linear regression analysis is based on zonal-mean tropopause height, the regressed fields may selectively highlight convective activity, which happens to excite global-scale responses. However, the artifact of regression seems to be minimal. Although not shown, almost identical results are reproduced when the UT temperature fields are regressed against a local temperature anomaly (or equivalently \(Z_T\) over A. The same is true for the OLR field in Fig. 6d when OLR anomalies are regressed against \(\text{OLR}^{\text{wp}}\). These findings suggest that the key patterns revealed in the regressed fields are the representative coherent structures in the tropical UT.

During the NH winter and SH spring, there is an apparent link between these tropical processes and high-latitude tropopause variability, mostly likely through the stratospheric circulation. While this phenomenon is interesting in its own right, the analyses presented in this study indicate that the impact of this remote link on the intraseasonal time-scale tropical tropopause variability is secondary. This contrasts the seasonal and interannual tropical tropopause variability, for which the stratospheric circulation is known to play the major role.

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REFERENCES


CORRIGENDUM

An error occurred on the title page of “Intraseasonal Variability of the Zonal-Mean Tropical Tropopause Height,” by Seok-Woo Son and Sukyoung Lee, which was published in the Journal of the Atmospheric Sciences, Vol. 64, No. 7, 2695–2706, resulting in an incorrect DOI number being printed.

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The staff of the Journal of the Atmospheric Sciences regrets any inconvenience this error may have caused.