Vertically Propagating Kelvin Waves and Tropical Tropopause Variability

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ABSTRACT

The relationship between local convection, vertically propagating Kelvin waves, and tropical tropopause height variability is examined. This study utilizes both simulations of a global primitive-equation model and global observational datasets. Regression analysis with the data shows that convection over the western tropical Pacific is followed by warming in the upper troposphere (UT) and cooling in lower stratosphere (LS) over most longitudes, which results in a lifting of the tropical tropopause. The model results reveal that these UT–LS temperature anomalies are closely associated with vertically propagating Kelvin waves, indicating that these Kelvin waves drive tropical tropopause undulations at intraseasonal time scales.

The model simulations further show that regardless of the longitudinal position of the imposed heating, the UT–LS Kelvin wave reaches its maximum amplitude over the western Pacific. This result, together with an analysis based on wave action conservation, is used to contend that the Kelvin wave amplification over the western Pacific should be attributed to the zonal variation of background zonal wind field, rather than to the proximity of the heating. The wave action conservation law is also used to offer an explanation as to why the vertically propagating Kelvin waves play the central role in driving tropical tropopause height undulations.

The zonal and vertical modulation of the Kelvin waves by the background flow may help explain the origin of the very cold air over the western tropical Pacific, which is known to cause freeze-drying of tropospheric air en route to the stratosphere.

1. Introduction

The established view is that the height of the time-mean tropical tropopause is controlled by deep convection up to a height of \( \approx 12–13 \) km (Riehl and Malkus 1958), and that radiative balance explains the continued decrease in temperature up to the cold-point tropopause at \( \approx 17 \) km. In contrast, the temporal variability of the tropical tropopause height, and related variables such as tropopause temperature, are known to be associated with fluctuations in upwelling induced by stratospheric extratropical wave breaking (Holton et al. 1995) and perhaps in the strength of the Hadley circulation (Plumb and Eluszkiewicz 1999).

At intraseasonal time scales, there is increasing evidence that tropical tropopause temperature fluctuations are associated with Kelvin waves. Zhou and Holton (2002) found that the cooling of the cold-point tropopause in the tropics is associated with the Madden–Julian oscillation (MJO; Madden and Julian 1971), and they attributed this cooling to the presence of Kelvin waves. Based on radiosonde data, collected over a period of 1 month over eastern Java, Indonesia, Tsuda et al. (1994) reported that the tropopause descends following the phase lines of 20-day Kelvin waves, which appear to be excited by tropical convection. With a linearized primitive-equation model Salby and Garcia (1987) and Garcia and Salby (1987) showed that localized tropical heating can generate similar vertically propagating Kelvin waves in the stratosphere. Salby and Garcia (1987) found that short-term heating fluctuations generate vertically propagating waves, which resemble observed Wallace and Kousky (1968)
Kelvin waves, while seasonal time-scale heating produces a Walker circulation–like Kelvin wave, which is confined to the troposphere. The relationship between the convection and Kelvin waves was observationally established by Takayabu (1994), Wheeler et al. (2000), Kiladis et al. (2001), and Randel and Wu (2005).

The above studies collectively suggest that convective heating, equatorial Kelvin waves, and tropical tropopause height undulations are closely related. In fact, with Challenging Minisatellite Payload (CHAMP)/global positioning system (GPS) radio occultation measurements collected between May 2001 and October 2005, Ratnam et al. (2006) showed that the seasonal and interannual variation in the zonal wavenumber 1 and 2 Kelvin wave amplitudes coincide with those of the tropical tropopause height. They also show that statistically significant correlations exist between these planetary-scale Kelvin wave amplitudes and outgoing longwave radiation (OLR) in the tropics. However, because each of these studies focused on either one or two of these processes (Zhou and Holton 2002; Wheeler et al. 2000; Kiladis et al. 2001), and because the time period of some of these analyses was limited (Tsuda et al. 1994; Randel and Wu 2005), it is still difficult to draw a general coherent picture of how each of these processes are related. While the results of Ratnam et al. (2006) establish the relationship between convection, Kelvin waves, and variability of tropical tropopause height, their focus was on seasonal and interannual time scales, not on the global-scale Kelvin wave and its relationship with intraseasonal tropopause height variations.

In a more recent study, Son and Lee (2007) showed that the lifting of the zonal mean tropical tropopause is preceded first by MJO heating over the Pacific warm pool, followed by the excitation of a large-scale circulation, which rapidly warms (cools) most of the tropical troposphere (the lowest few kilometers of the tropical lower stratosphere). However, they could not discern whether the lower stratosphere (LS) cooling is associated with vertically propagating Kelvin waves (Tsuda et al. 1994; Randel and Wu 2005), or if it reflects thermal wind adjustment in response to Gill-type Kelvin and Rossby waves (Highwood and Hoskins 1998). In addition, because the analysis of Son and Lee (2007) was based on the height of the zonal mean tropical tropopause, their analysis may have selectively highlighted convective activity, which excites a global-scale response.

In this study, we investigate the relationship between convective heating, vertically propagating equatorial Kelvin waves, and the tropical tropopause height. As will be shown in this study, the LS Kelvin wave is a robust response to localized tropical heating, and the Kelvin wave response is greatest in the western tropical Pacific. By presenting a plausible mechanism that can account for the Kelvin wave’s amplitude modulation in both the zonal and vertical directions, this study also attempts to provide an explanation for why the Kelvin wave response is greatest in the western and central tropical Pacific, and why vertically propagating Kelvin waves play a pivotal role for tropical tropopause undulations.

The primary tool in this study is a multilevel primitive-equation model on the sphere. Observational analyses are also presented. Section 2 briefly describes the data and the model. Results from a regression analysis with a global dataset will be presented in section 3, and the model simulations will be described in section 4. Section 5 presents a wave action analysis on the effect of the background flow on vertically propagating Kelvin waves. This will be followed by the summary and conclusions in section 6.

2. Data and model

a. Data and analysis method

The data and regression methods are identical to those used in Son and Lee (2007). The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset is utilized. Daily data on sigma levels, which span the time period from 1 January 1979 to 31 December 2004, are interpolated with a cubic spline onto height coordinates with a 0.5-km interval. The tropopause height is then identified with the standard lapse-rate criterion—the lowest level at which the temperature lapse rate remains below 2 K km⁻¹ for a vertical distance of 2 km. The tropical tropopause identified by this procedure compares very well with the tropical tropopause pressure provided by NCEP–NCAR (not shown). The intensity of the convection is estimated from the daily OLR dataset, archived by the National Oceanic and Atmospheric Administration (NOAA).

In all datasets, the seasonal cycle, which is defined by smoothed, calendar-day mean values, is removed. Interannual variability is also removed by subtracting the seasonal mean for each year. To more precisely describe these operations, we write the variable of interest, say $T$, in the following manner:

$$T(i, j) = T_s(i) + T_x(j) + T'(i, j).$$

The first term on the right-hand side, $T_s(i)$, is the smoothed calendar mean, defined as $\mathcal{L}[(1/N)\sum_{j=1}^{N} T(i, j)]$, where $\mathcal{L}$ is a low-pass filter with a 20-day cutoff, $N$
the total number of years, \( i \) the date within each year, and \( j \) the year index. The second term, \( T_s(j) \), is the seasonal mean for each year after the seasonal cycle has been removed, \( T_s(j) = \frac{1}{M} \sum_{i=1}^{M} [T(i, j) - T_s(\delta)] \), where \( i = 1 \) corresponds to 1 December, and \( i = M \) to 28 February. The filtered temperature anomalies are denoted by \( T' \) and the filtered tropical tropopause height anomalies by \( Z' \), where the prime denotes a deviation from the seasonal cycle and seasonal mean for each year. These anomalies are averaged from 10°S to 10°N, which will be indicated with an angled bracket.

The above tropical mean anomalies are regressed against the OLR anomalies of the western Pacific, which are defined to span the region between 120°E and the date line, and from 10°S to 10°N (hereafter denoted as \( \text{OLR}_{\text{wp}} \)). Smoothing is not applied, and the resulting values are set to correspond to a one standard deviation value of \( \text{OLR}_{\text{wp}} \). We consider only the Northern Hemisphere winter, when the MJO convection is strongest [see Son and Lee (2007) for a discussion on the seasonal differences].

**b. Model and experiment design**

The numerical model used in this study is the dry dynamical core of the Geophysical Fluid Dynamic Laboratory (GFDL) global spectral model (Feldstein 1994; Kim and Lee 2001; Son and Lee 2005). Rhomboidal 30 resolution is used and there is no topography. The vertical levels are defined in \( \sigma \) coordinates. There are 60 levels, starting from 0.975 near the surface up to 0.0012 (about 45 km at the equator). To accurately resolve upper-troposphere (UT)–LS features, fine vertical resolution, corresponding to about 90 m, is used near the tropopause (see Fig. 1b, which shows the vertical heating profile). Throughout the troposphere, the vertical resolution is approximately 1 km.

Two different basic states are used. One state is a zonally asymmetric (3D) and the other is zonally symmetric (2D). The former state is the December–February (DJF) time mean flow for the years 1948 through 1996, and the latter state is the corresponding zonal mean. Because the climatological 3D state is not balanced, a forcing term is added to ensure that the 3D state is a solution of the model equations (e.g., James et al. 1994; Franzke et al. 2004). Franzke et al. (2004) discuss the limitations of this approach and provide a rough estimate for the length of the integration period for which the basic state can be regarded as a steady solution. The length of this integration is inversely proportional to the strength of the climatological eddy fluxes, where eddy is defined as the deviation from the climatological flow. According to their estimation, this period lasts for about 20 days in the North Atlantic.

While this time scale is expected to be greater in the tropics, we take the 20-day limit as rough guidance for our model integration.

Newtonian cooling is applied to the perturbation temperature, where a perturbation is defined as the deviation from the basic state. The time scale of the Newtonian cooling is 5 days in the stratosphere and 20 days in the middle and upper troposphere (cf. Jin and Hoskins 1995; Taguchi 2003). Throughout the lower troposphere (\( \sigma \approx 0.725 \)), the Newtonian cooling time scale gradually decreases to 8 days at the lowest model level. There is also friction at the lowest four model levels (\( \sigma \approx 0.825 \)), with the time scale decreasing from 4.5 to 1.5 days at the lowest level. The surface value of the vertical diffusion coefficient \( \nu \) is set to 0.5 \( \text{m}^2\text{s}^{-1} \) and the value of \( \nu \) is chosen to keep the dynamic dif-

![Fig. 1. (a) The horizontal structure of the idealized heating at 400 hPa. The contour interval is 1 K day\(^{-1}\). (b) The vertical profile of the idealized heating at 150°E and the equator. Open circles indicate the vertical grid points used for the model.](image-url)
fusion coefficient $\nu$, constant with height, where $\rho$ is density. To prevent contamination by vertical wave reflection, we apply very large values of the diffusion coefficient to the highest eight levels ($\sigma = 0.01$). For the horizontal diffusion, an eighth-order diffusion scheme is used with a horizontal diffusion coefficient of $8 \times 10^{-7}$ m$^2$ s$^{-1}$.

For the diabatic heating, an elliptic horizontal distribution is utilized, which is similar to the intertropical convergence zone (ITCZ) proxy used by Nieto Ferreira and Schubert (1997). The heating field at 400 hPa, which is centered at 150°E and the equator, is shown in Fig. 1a. The zonal and meridional extents of the heating are 100° and 10°, respectively. The vertical profile of the heating takes the form of two parabolas joined at 400 hPa, as shown in Fig. 1b. It has a maximum value of 4.8 K day$^{-1}$ at 400 hPa and decreases to 0 at 975 and 175 hPa. The heating is gradually turned on (off) during the 1st (10th) day. The response of the model atmosphere to the idealized heating is defined as the deviation from the basic state prescribed at the initial time.

3. Observations

a. Tropical tropopause fluctuations

Figure 2 shows the temporal evolution of $\langle Z_T \rangle'$ (the main panel) regressed against the convective activity over the Pacific warm pool, measured by a negative one standard deviation in $\langle OLR^{wp} \rangle$ at lag 0. As indicated in section 2, the angled bracket denotes a latitudinal average from 10°S to 10°N. The positive (negative) time lag denotes the days after (before) the minimum daily $\langle OLR^{wp} \rangle$ value. Superimposed upon the contours in Fig. 2 is the $\langle OLR \rangle'$ field regressed against $\langle OLR^{wp} \rangle$. To indicate the location of the convection, values less than $-2$ W m$^{-2}$ are shaded. It is worth noting that while
no filtering is performed other than the removal of the seasonal cycle and the interannual averages, the regressed (OLR)' field shows eastward propagation at the MJO time scale. In fact, Son and Lee (2007) performed spectral analysis on the daily (OLRwp)', and found a spectral peak at 45 days, which is statistically significant above the 95% confidence level against the red-noise null hypothesis.

The autoregression of (OLRwp)' is shown with shading in the right panel of Fig. 2. The maximum value of the zonal mean (ZT)' occurs at a lag of 7–8 days (see the right panel in Fig. 2). Because the value of zonal mean (ZT)' is close to zero at lag −2 days, it takes about 10 days for the zonal mean (ZT)' to reach its maximum value. At the time of the maximum zonal mean (ZT)', a positive (ZT)' contribution comes from most longitudes. This tropics-wide contribution results from an eastward propagation of (ZT)', as indicated by the thick dashed line, which is an approximate linear fit of the longitude and the time at which the (ZT)' value first exceeds 50 m. The local maxima in (ZT)' are statistically significant above the 99% confidence level. In addition, similar results are obtained when the regressions are performed against the NCEP–NCAR tropopause pressure (not shown).

b. Temperature perturbations

The results described above indicate that (i) the western Pacific convective heating typically gives rise to a lifting of the zonal mean (ZT)', which takes place over a 10-day time period; (ii) fluctuations in (ZT)' over the western Pacific play a key role for modulating the zonal mean (ZT)' response; and (iii) the contribution to the positive zonal mean (ZT)' comes from most longitudes.

The dynamical link between the western Pacific heating and the tropopause height can be grasped in a more coherent manner by examining the associated temperature anomalies, because the tropopause is typically raised by UT warming–LS cooling. Figure 3 shows the anomalous temperature (T)' and wind vector field ((u)', (w)'), regressed against (OLRwp)'). The vertical component of the wind (w)' is scaled by multiplying by a factor of 1440, so that the slope of the wind vector is consistent with the aspect ratio of the zonal and vertical axes. As expected, based on the location of the convective heat-
ing, the maximum \(\langle T \rangle \) occurs in the midtroposphere over the Pacific warm pool. In addition to this tropospheric warming, there is also a wavy pattern in \(\langle T \rangle \) in the vicinity of the tropopause, which bears resemblance to Fig. 6 of Kiladis et al. (2005), who display temperature fields regressed against an MJO index. The MJO index in Kiladis et al. (2005) is defined by the eastward-propagating component of the MJO time-scale OLR anomalies. Given that our index (OLR\(^{mp}\)) has a spectral peak at the MJO time scale (Son and Lee 2007), it is not surprising that the two temperature fields—one regressed against (OLR\(^{mp}\)) and the other against an MJO index—resemble each other.

The lack of vertical resolution in the reanalysis data in the LS region makes it difficult to identify the planetary-scale wavy features seen in the UT–LS region. However, given the observational and theoretical expectations that the LS region is dominated by vertically propagating planetary-scale Kelvin waves (Wallace and Kousky 1968) and mixed Rossby–gravity waves (Yanai and Maruyama 1966), the UT–LS features in Fig. 3 appear to be accounted for, in part, by the planetary-scale vertically propagating Kelvin waves. A visual inspection of Fig. 3 indicates that the zonal scale is dominated by zonal wavenumbers 1–2, the vertical wavelength in the LS region is about 6 km, and the wavy features are tilted eastward with height. These characteristics are consistent with vertically propagating Kelvin waves, not with the mixed Rossby–gravity waves (see Table 4.1 in Andrews et al. (1987) for a summary of the wave characteristics).

For the sake of clarity, it needs to be mentioned that although excited by convective heating, vertically propagating LS Kelvin waves in the tropics are not coupled to convection. Chang (1976) demonstrated that randomly distributed convective heating can generate vertically propagating Kelvin waves, and above the forcing region Kelvin waves of the largest zonal scale are most effectively excited. Because of the absence of coupling, Randel and Wu (2005) described the LS vertically propagating Kelvin waves as “free” waves, as opposed to convectively coupled waves (Takayabu 1994; Wheeler and Kiladis 1999; Wheeler et al. 2000).

4. Numerical model experiments

The key features highlighted in the previous section will be examined with an idealized model. Because the vertical resolution in the NCEP–NCAR reanalysis is very low above the troposphere, regardless of the reliability of the analysis, the representation of the LS features is not adequate. However, the analysis in the previous section indicates that the LS features are critical for tropical tropopause height fluctuations. While a comparison with direct observations, such as radiosonde data (Kousky and Wallace 1971; Tsuda et al. 1994, etc.), is the most persuasive means to verify the features revealed by the regression analysis, availability of such direct observations is limited both in time and space. Although simulations from a numerical model cannot be used to verify the reanalysis data, with a sufficiently high resolution in the vertical direction, the model simulation can help reveal the LS features, which are not well represented in the reanalysis. More importantly, as will be described in section 4b, the model can be used to investigate the mechanism behind UT–LS Kelvin wave modulation.

a. Western Pacific heating

In the simulation to be presented in this section, the idealized heat source (see section 2b) is placed in the western equatorial Pacific in order to model the convective heating indicated by the regressed OLR field. To mimic the eastward propagation of the OLR field, the center of the heating is prescribed to stay at 150°E for the first 4 days, and then to move eastward at a speed of 30° (5 days)\(^{-1}\).

1) Model-simulated Kelvin wave structure

Prior to presenting the model simulation of the tropical tropopause undulation, we examine the three-dimensional structure of the model response both for comparison with those results obtained from previous studies, and to help understand the model’s tropopause response to the heating. For ease of comparison with the previous studies (Gill 1980; Jin and Hoskins 1995), we first show the results with the 2D basic state. The left column in Fig. 4 displays the UT perturbation pressure and horizontal wind fields. A Gill-type response is evident, with the anticyclonic Rossby wave dipole to the west of the heating, and the Kelvin wave to the east of the heating. The Rossby wave signal is stronger in the Northern Hemisphere because of the asymmetry of the background flow about the equator. The Kelvin wave can be identified by the equatorial westerlies. The leading edge of the westerlies advances to 60°W at day 5, to the Greenwich, United Kingdom, meridian by day 7, and to 30°E by day 9. This eastward expansion of the UT equatorial westerlies (seen in Fig. 4) can be inferred in the streamfunction field shown in Fig. 4a of Jin and Hoskins (1995). Likewise, the UT wind fields, simulated with the 3D basic state (the left column in Fig. 5), are also consistent with the streamfunction field shown by Jin and Hoskins (cf. day 9 in Fig. 5 with their Fig. 12a). Although there are differences in both the 2D and the 3D model runs, the UT Kelvin waves are anchored...
to the local heating, as for the time mean Walker circulation (e.g., Gill 1980; Salby and Garcia 1987; Jin and Hoskins 1995).

In contrast to the above UT response, the LS Kelvin wave response is not anchored to the heat source. At the 20-km level, a patch of equatorial westerlies (the right column in Figs. 4 and 5) travels eastward and encircles the globe in 10 days. Figure 6 shows the UT–LS vertical cross section of the perturbation pressure (shaded), temperature (contours), and wind vectors from the 3D simulation. (Having demonstrated that the 2D simulation conforms to that of previous studies, and

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**Fig. 4.** The perturbation pressure (contours) and wind (arrows) at (left) 14 and (right) 20 km simulated with the 2D basic state for (top to bottom) days 3, 5, 7, 9. The contour interval is (left) 10 and (right) 2 Pa. The location of the heating is indicated with shading. Wind vectors are in m s$^{-1}$. 

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because the essential feature of the equatorially trapped Kelvin wave structure is also faithfully simulated in the 3D run, for the sake of brevity, the vertical cross section from the 2D simulation is not shown.) The phase relationships between the pressure and wind fields, and between the wind and temperature fields, are consistent with the theoretical phase relationships of a vertically propagating Kelvin wave (Wallace and Kousky 1968). As can be seen, the westerly perturbations are in phase with high pressure perturbations, consistent with the fact that the equatorial Kelvin wave is in geostrophic balance. As expected by the fact that the Kelvin wave is driven by buoyancy, the upward motion leads the negative temperature perturbation by
¼ of a cycle. This phase relationship can be seen by comparing the day-5 wind with the day-7 temperature field, and the day-7 wind field with day-9 temperature field. As expected for a Kelvin wave, which is excited from below, downward phase propagation can be seen.

2) TROPICAL TROPAUSE UNDULATION

To investigate how this Kelvin wave response is associated with the tropopause undulation, Fig. 7 recasts Fig. 6, ranging from 5 to 25 km, showing simulations out to 17 days. Simulations from days 9 to 17 show that the vertical scale of the LS Kelvin wave is about 6–10 km and the period is about 10–15 days, further conforming to the observed Wallace–Kousky Kelvin wave. More importantly for the tropopause undulation, it can be seen that when the heating is turned on (recall that the heating is turned off on day 10), a shallow layer of cooling occurs at the tropopause level directly above the heating. In particular, the day-5 temperature response resembles the buoyancy solution of Pandya et al. (1993) for an idealized heating in a Boussinesq fluid. For all the tropical locations and time scales that they...
Fig. 7. Longitude–height cross section of perturbation temperature (K, contours), wind vectors (m s\(^{-1}\), arrows), and tropopause height (scale is indicated on the right side), generated with the 3D basic state for (top to bottom) days 3–17. The thick solid line indicates the DJF climatological tropopause. All quantities shown are averaged between 10°S and 10°N. The vertical wind is multiplied by 1.44 \times 10^3 so that the slope of the wind vectors is consistent with the aspect ratio of the figure. The location of heating is indicated with a horizontal bar under each frame.
examined, Holloway and Neelin (2007) found with satellite and reanalysis data that a shallow layer of tropopause-level cooling occurs in conjunction with tropospheric heating. Using a Boussinesq equation set, they provided an explanation as to why a shallow layer of cooling must occur above a tropospheric heat source. Because their explanation describes how a heat source generates internal gravity waves in a hydrostatic atmosphere, the same explanation can be given for the generation of tropopause-level cooling in our simulations.

In the simulation presented in this study, the tropopause-level cooling explained above plays the key role for the tropopause height variability. Figure 7 shows that the tropopause rises (dotted lines indicate the deviation from the time mean tropopause) in regions where the cooling occurs just above the time mean tropopause (thick horizontal lines). Because the Kelvin wave response occurs to the east of the heat source, the center of the cooling does not occur directly above the heating region, but instead is shifted to the east. In addition, because the phase lines for vertically propagating Kelvin waves tilt eastward with height, cooling occurs just above (below) the tropopause in the eastern (western) part of the heating. This causes the tropopause to ascend in the eastern side of the heating, and to descend in the western side. This descent is further aided by warming, which begins to appear on day 5 just above and to the west of the cooling. In the same region, the wind field on day 3 (see Fig. 6) shows that this warming is preceded by sinking motion, consistent with the earlier analysis that the LS response represents a vertically propagating Kelvin wave.

To the east of the heating, the LS cooling and the attendant elevated tropopause rapidly propagate to the east. Referring back to Fig. 5 and the foregoing discussion, we recognize that this eastward propagation reflects the LS Kelvin wave. In Fig. 8, this eastward propagation of the elevated tropopause is indicated with a thick dashed line. It can be seen that it takes about 10 days for the elevated tropopause to encircle the entire tropics. During the eastward expansion of the elevated tropopause, the height of the zonal mean tropopause gradually increases (right panel of Fig. 8). However, because the LS cooling is no longer forced...
after the heat source is turned off at day 10 (notice that the LS cold anomaly no longer exist above the heating on day 11), the zonal mean tropopause height declines after day 10.

With the exception of the negative perturbation near 60°W on days 12 and 13, the regressed and simulated tropopause height perturbations resemble each other (cf. Fig. 2 with Fig. 8, while noting that lag- N day in Fig. 2 should be compared with a model day, which is greater than N, say N + 5, in Fig. 8). This is because the lag-0 day corresponds to the time of the maximum OLR, meaning that the OLR anomaly amplitude is still substantial at small ± lag days.) In addition, in the immediate vicinity of the tropopause, below 20 km, the regressed LS temperature field (Fig. 3) compares well with the simulated LS temperature field (Fig. 7).

b. Sensitivity to the location of heat source

The foregoing analysis indicates that the simulated tropopause undergoes undulation as the LS Kelvin wave propagates both vertically and zonally. In addition, the LS Kelvin wave amplitude is greatest over the western Pacific. This large Kelvin wave amplitude over the western Pacific may be attributed to the proximity to the heating. However, the subsequent evolution from day 13 to 15 shows that as the Kelvin wave packet reenters the western Pacific, the LS cold anomaly in that region strengthens once more, despite the fact that the heating has been turned off 5 days earlier. These findings raise the interesting possibility that the zonal and vertical variation of the LS Kelvin wave structure may hinge on the ambient background flow, more so than on the proximity to the heat source. This possibility will be explored further in sections 4b and 5.

To test the idea that the LS Kelvin wave structure may be influenced more directly by the background flow than by the proximity of the heat source itself, additional simulations are performed. In one simulation, the heat source is prescribed in the eastern Pacific centered at 120°W, but the remaining model setup is exactly the same as before. Figure 9 shows the result, which is analogous to Fig. 7. Unlike the run with western Pacific heating, the LS Kelvin wave response, if any, is very weak directly above the heating. The LS Kelvin wave signal does not become apparent until the Kelvin wave packet nears the Greenwich meridian. As the wave packet enters the Indian Ocean and western Pacific, the wave amplifies and the wave signal extends further into the stratosphere (days 13–17). The overall Kelvin wave structure in the stratosphere during days 15–17 is remarkably similar to that on day 9 of the western Pacific heating simulation, with largest amplitude over the western Pacific.

Although not shown, five additional experiments are performed, with the prescribed heat source centered at the Greenwich meridian, 60°E, 120°E, 180°, and 60°W. In all of these simulations, the strongest LS Kelvin wave response once again occurs over the western Pacific. This result demonstrates that the background flow over the western Pacific is responsible for the large Kelvin wave signal in that region. This, however, is not to say that the western Pacific heating is irrelevant for the Kelvin wave modulation. Because the characteristics of the background flow must be determined in part by tropical convective heating, the western Pacific heating must influence the Kelvin wave structure, but apparently its impact is made indirectly through the background flow.

c. Zonally symmetric basic state

The impact of the zonal modulation of the Kelvin wave on the zonal mean tropopause undulation is tested with a third simulation in which the basic state is the zonal mean of the DJF climatological state. As shown in Fig. 10, there is very little zonal variation in the Kelvin wave amplitude. Once again, the tropopause rises as cold Kelvin wave perturbations position themselves just above the tropopause (not shown). However, the maximum zonal mean tropopause height perturbation in this simulation is 38 m (not shown), about 15% smaller than those from the two previous simulations with a zonally varying basic state. This indicates that, all else being equal, the zonal variation of the background flow enables the tropopause height to undulate by a greater margin.

5. Influence of background state on Kelvin waves

The foregoing model simulations indicate that, regardless of the geographical location of the imposed tropical heating, the temperature perturbation associated with the LS Kelvin wave is greatest over the western Pacific. This western Pacific amplification of the Kelvin wave signal may also be highly relevant for the issue of stratosphere–troposphere exchange (STE), namely, the origin of the cold region, known as the “cold trap.” According to the “stratospheric fountain” hypothesis of Newell and Gould-Stewart (1981), tropospheric air enters the stratosphere through the cold trap, which is a relatively small, cold, confined region over the western tropical Pacific. The temperature in the cold trap is so low that the moisture within the tropospheric air “freeze dries.” This leaves behind very dry air that then enters the stratosphere. Various mechanisms have been suggested to explain this cold
FIG. 9. As in Fig. 7, but the heating is centered at 120°W.
Fig. 10. As in Fig. 7, but the zonally averaged DJF flow is used as the basic state.
trap, including convective detrainment (Sherwood and Dessler 2001; Sherwood et al. 2003; Kuang and Bretherton 2004) and a hydrostatic gravity wave response (Holloway and Neelin 2007) to a heat source.

Tsuda et al. (1994) and Randel and Wu (2005) recognized that the LS cold perturbations associated with the Kelvin wave may be an answer to this question. Because the Kelvin wave is one particular type of gravity wave, the Kelvin wave mechanism and the mechanism of Holloway and Neelin (2007) are not mutually exclusive. An important difference, though, is that while the Holloway–Neelin mechanism explains a layer of cooling directly above the heat source, a near-field response, the Kelvin wave mechanism and the mechanism of Holloway and Neelin (2007) are not mutually exclusive. An important difference, though, is that while the Holloway–Neelin mechanism explains a layer of cooling directly above the heat source, a near-field response, the Kelvin wave mechanism and the mechanism of Holloway and Neelin (2007) are not mutually exclusive.

With these questions in mind, in this section, we appeal to linear wave theory to gain insight into how the UT–LS Kelvin wave amplitude varies with the background state. To do so, we consider the dispersion relation for a hydrostatic Kelvin wave, which takes the form of

\[ \omega = U k \pm N k / m, \]  

where \( U \) is the background zonal wind, \( N \) the background buoyancy frequency, \( \omega \) the ground-based frequency, and \( k \) and \( m \) the zonal and vertical wavenumbers, respectively. Taking the sign of \( \omega \) as positive definite, a positive value of \( k \) represents eastward phase propagation, and vice versa. Similarly, a positive (negative) value of \( m \) indicates upward (downward) phase propagation. With this sign convention, \( k > 0 \) for equatorially trapped Kelvin waves. Because the Kelvin wave in question is excited from below, the vertical component of the group velocity must be positive. This constraint requires that \( m < 0 \), which results in the negative root in (1) being the only valid dispersion relation (Holton 1992; Andrews et al. 1987). It follows then that

\[ c - U > 0, \]  

for Kelvin waves, where \( c \) is phase speed.

To examine how the Kelvin wave amplitude modulation depends upon the zonal and vertical variation of the background state, we consider the wave action conservation law (Lighthill 1978). For equatorially trapped waves (Andrews and McIntyre 1976), wave action conservation takes the form of

\[
\frac{\partial A}{\partial T} + \frac{\partial}{\partial X} (AC_{gx}) + \frac{\partial}{\partial Z} (AC_{gz}) = 0,
\]

following a ray, which is the path parallel to the local group velocity, \( C_{gx}(x, z, T) \) and \( C_{gz}(x, z, T) \), where \( T \) and \( x, z \) are the slowly varying time and spatial scales, as required by the Wentzel–Kramers–Brillouin (WKB) approximation \( T = \epsilon^{-1}(2\pi/\omega) \), \( x = \epsilon^{-1}(2\pi/k) \), and \( z = \epsilon^{-1}(2\pi/m) \), where \( \epsilon \) is a small parameter (Andrews et al. 1987). For the wave field in question,

\[
C_{gx}(x, z, T) = U(x, z) - N(x, z)m(x, z, T), \tag{4}
\]

and the wave action \( A = \langle E \rangle \omega_{\epsilon}^{-1} \), where \( \omega_{\epsilon} \) is the relative frequency and \( \langle E \rangle \) is the meridionally integrated wave energy density (in units of wave energy per area), that is,

\[
\langle E \rangle = \int_{-\infty}^{\infty} \frac{\rho_{\epsilon}}{2} (u^{2} + N^{-2} \phi^{-2}) \, dy, \tag{6}
\]

in log-pressure coordinates (Andrews et al. 1987). Therefore, the use of (3), the wave action conservation law, allows us to determine how \( \langle E \rangle \), which we use as a measure of the Kelvin wave amplitude, depends upon the background state. In (6), \( u \) is the zonal wind and \( \phi \) is the geopotential associated with the wave, \( \rho_{\epsilon} \) is the basic-state density, and \( \phi_{z} = RH^{-1}T \), where \( R \) is the gas constant and \( H \) the density-scale height. In practice, the meridional integral in (6) should be taken over finite limits. Given that the Kelvin wave decays exponentially in the meridional direction, a natural choice for these limits is the \( e \)-folding distance (separately determined for the Northern and Southern Hemispheres) from the equator. In calculating (6), we used the \( e \)-folding distance at each longitude for both hemispheres and for each model day. The resulting values were averaged from day 7 to 17, and from 17 to 25 km. It turned out that zonal variation in the \( e \)-folding distance was small. As a result, the temperature and wind fields, integrated over this zonally varying \( e \)-folding distance (not shown), were found to be very similar to those obtained by integrating from 10\( ^{\circ} \)S to 10\( ^{\circ} \)N, as in Figs. 7 and 9.

In principle, given \( U \) and \( N \), and the initial values for \( A, k, \) and \( m, A(x, z, T) \) can be solved by integrating (3) and the following two equations:

\[
d_{x}k(x, z, T)/dt = -\omega A(x, z, T)/\partial X,
\]

\[
d_{x}m(x, z, T)/dt = -\omega A(x, z, T)/\partial Z,
\]
where $\frac{dA}{dt}$ denotes the rate of change following the local group velocity (Andrews et al. 1987). However, to gain physical insight into how $A$ (and therefore $E$) is influenced by $U(x, 2)$ and $N(x, 2)$, we take an alternative approach and consider an approximate wave conservation equation for which there is an analytical expression for $A$. To do so, we assume that the fractional variation of $k$ and $m$ with respect to $U$ and $N$ is much less than 1, and therefore that $k$ and $m$ can be treated as constants. The fully developed stratospheric Kelvin wave structure, shown in the right panel of Fig. 7, indicates that this is a reasonable approximation, because $k$ and $m$ do not vary by more than a factor of 2, while variations in both $U$ and $N$ ($U$ is shown in Fig. 11, and $N$ can be inferred from Fig. 12a) are much greater.

To determine which term in (3) dominates, we calculate the slope of the ray [see (4) and (5)],

$$S = \frac{C_{gx}}{C_{gz}} = \frac{N(x, z)k_0/m_{o}^2}{U(x, z) - N(x, z)/m_{o}},$$

where $x$ and $z$ are used in place of $X$ and $Z$, respectively, to indicate that observed values of $U$ and $N$ are used, and the subscript $o$ emphasizes that $k$ and $m$ are treated as constants.

Figure 12 shows an estimation of $C_{gz}$ and $C_{gx}$ for the Kelvin wave and the slope of the ray $S$ at 3.35°N. For regions where $|C_{gx}| \leq 10^{-3} \text{m s}^{-1}$, contours of $S$ are not drawn. This latitude is chosen because the largest value of $S$, excluding the regions where $|C_{gx}| \leq 10^{-3} \text{m s}^{-1}$, is found at 3.35°N. The values of $k_0$ and $m_0$, used for the estimation in Fig. 12 are $1.57 \times 10^{-7} \text{m}^{-1}$ and $-6.28 \times 10^{-4} \text{m}^{-1}$, respectively, corresponding to a zonal wavenumber 1 Kelvin wave with a vertical scale of 10 km. The structure of $C_{gz}$ largely reflects the static stability, which has both large values just above the tropopause and little zonal variation (Fig. 12a). In contrast, the zonal variation of $C_{gx}$ is much greater, with large values over the eastern Pacific where $U > 0$ and small values over the western Pacific where $U < 0$ (cf. Fig. 12b with Fig. 11). As a result, the zonal variation of the slope $S$ is determined mostly by $C_{gx}$, with the steepest slope occurring over the western Pacific (Fig. 12c). Except in the region where the value of $U$ approaches $-10 \text{m s}^{-1}$, the slope $S$ ranges from $10^{-4}$ to $10^{-3}$ (Fig. 12c).

### a. Horizontal structure

For a conservative steady wave, the first term in (3) is zero and, to determine whether the primary balance is achieved by the horizontal flux divergence term, $C_{gx}(\partial A/\partial X) + A(\partial C_{gx}/\partial X)$, or by the vertical flux divergence term, $C_{gz}(\partial A/\partial Z) + A(\partial C_{gz}/\partial Z)$, we perform the following scaling analysis. By assuming that the horizontal scale of $A$ and $C_{gx}$ are equal, we write the horizontal scale $L$ such that $L = \left(|(1/A)(\partial A/\partial X)|^{-1} = |(1/C_{gx})(\partial C_{gx}/\partial X)|^{-1}\right.$. Making a parallel assumption for the vertical scale, we write the vertical scale $D = \left(|(1/A)(\partial A/\partial Z)|^{-1} = |(1/C_{gz})(\partial C_{gz}/\partial Z)|^{-1}\right.$.

It follows that $O[C_{gx}(\partial A/\partial X)] = O[A(\partial C_{gx}/\partial X)] = AC_{gx}/L$, and
fore, if

\[ O(C_{gz} / C_{gx}) \approx 3 \times 10^{-4} \] (see Fig. 12c), \( O(L) \approx 1 \times 10^7 \) m (Fig. 12b), and \( O(D) \approx 1 \times 10^4 \) m, yielding \( O[(C_{gz} / C_{gx})(L/D)] \approx 3 \times 10^{-1} \). Thus, except for the limited region over the western Pacific (the region not contoured in Fig. 12c),

\[ C_{gx} A = (U - N/m \rho) [-(m/\rho)(Nk_o)] E \]

\[ = c k_o^{-1} (c - U)^{-1} E = \text{constant along } x. \] (10)

This relation reveals that if \( U < 0 \), \( (c - U)^{-1} \) must be small and positive [see (2)], and therefore the value of \( E \) must be large. Likewise, \( E \) must take on a smaller value in regions where \( 0 < U < c \).

If one relaxes the assumption of \( k \) being constant and allows \( k \) to evolve following the ray-tracing equation
d\( k \), \( k(X, Z, T)/dX = -\omega \), \( (\omega(X, Z, T)) \) \( dX \) under condition (8), it can be shown that

\[ k^2(X) = k^2(X) \left[ \frac{U(X) - N}{m \rho} \right] \]

where \( X \) is a reference longitude. Provided that the ratio \( N/m \rho \) does not vary with \( X \), this relation implies that \( k(X) \) is small (large) in the eastern (western) Pacific where \( U(X) \) is positive (negative). Because (10) indicates that small (large) \( k \) makes \( E \) even smaller (larger), allowing \( k \) to vary in \( X \) further strengthens the conclusion that \( E \) amplifies over the western Pacific where \( U < 0 \).

Because the basic-state density \( \rho_o \) and the buoyancy frequency \( N \) vary by a small amount in the zonal direction, variation in \( E \) mostly reflects variation in \( |u| \) and in \( |\phi_r| \) (or in \( T \) from the relation \( \phi_r = RH^{-1} T \)). Therefore, to the extent that (10) can be applicable, assuming that the partition between wave kinetic energy and wave available potential energy is constant, both the temperature and the zonal wind fields of the wave are expected to be proportional to \( (c - U)^{1/2} \) in the zonal direction. This prediction is indeed consistent with the simulations. In the western1 (eastern) Pacific where \( U < 0 \) (\( U > 0 \)), both temperature and wind perturbations are relatively large (small). This predicted amplification of the Kelvin wave in the westerlies is also consistent with the observational findings of Nishi and Sumi (1995) and Nishi et al. (2007). They found that because of the Wallace–Kousky Kelvin wave, the zonal wind near the tropical tropopause un-

\[ ^1 \text{Recall, however, that this analysis is invalid for the region between } 110^\circ \text{ and } 165^\circ \text{E.} \]
dynamics undergo rapid fluctuations, and the magnitude of this zonal wind fluctuation is greatest in the Eastern Hemisphere where the climatological zonal wind is easterly.

b. Vertical structure

In a manner parallel to the previous subsection, in regions where the rays are sufficiently steep so that

$$\frac{C_{gx} L}{C_{gx} D} \gg 1,$$

we may approximate the wave action conservation law as

$$C_{gx} A \approx \text{constant along } z.$$ (12)

Between $110^\circ$ and $165^\circ$E and between 10 and 17 km, Fig. 12c indicates that $O(C_{gx}/C_{gx}) \approx 10^{-1}$. Therefore, for $O(L) \approx 1 \times 10^7$ m and $O(D) \approx 1 \times 10^5$ m, we find that $O[(C_{gx}/C_{gx}(L/D))] \approx 10^2$, satisfying the inequality condition for the above equation. Neglecting the wave kinetic energy, the conservation law can be further approximated as

$$C_{gx} \rho_o \kappa^2 \approx \text{constant along } z,$$ (13)

where $\rho_o N^{-2} \phi_z^2$ is the wave potential energy per unit volume. As mentioned in the previous subsection, because $\phi_z$ is proportional to $T$, we treat $\phi_z$ as temperature. Substituting $Nk_o$ for $C_{gx}$, $m_{1/2}(Nk_o)$, and $\alpha_T^{-1}$, and using the dispersion relation $c - U = -N/m_{1/2}$, the left hand side of (12) can be rewritten as $(c - U)\rho_o N^{-3} \phi_z^2$. Thus, (12) implies that

$$|\phi_z(z)|$$

$$\approx \rho_o^{-1/2}(z) \rho_o^{-1/2}(z)\left[\frac{c - U(z)}{c - U(z)}\right]^{3/2} N(z) \left[\frac{N(z)}{N(z)}\right]^{3/2} |\phi_z(z)|$$

$$= \rho_o^{2/2N} U(z) \left[\frac{c - U(z)}{c - U(z)}\right]^{-1/2} N(z) \left[\frac{N(z)}{N(z)}\right]^{3/2} |\phi_z(z)|,$$ (13)

with the understanding that $c - U(z) > 0$ [cf. (2)], with $z_r$ being a reference height. The relation (13) predicts that away from the region of the forcing, in the absence of dissipation, and within a depth of less than twice the density-scale height, the Kelvin wave temperature is modulated by the background buoyancy frequency $N(z)$ to the power of $3/2$ and by the Doppler-shifted phase speed $c - U(z)$ to the power of $-1/2$. Because $N$ rapidly increases above the tropopause and reaches a local maximum near 20 km, and because the temperature depends on $N$ to a higher power than on $c - U(z)$, Kelvin wave temperature fluctuations are expected to be greatest in the vicinity of ~20 km, and to depend weakly on the background wind speed $U(z)$. Figures 13a–c display the height dependency of the three factors on the right-hand side of (13): $e^{(z-z_r)/2N}$, $[(c - U(z))/[c - U(z)]^{-1/2}$, and $[N(z)/N(z)]^{3/2}$. To be consistent with Fig. 12, climatological data at 3.35$^\circ$N are used for this evaluation. The value for $H$ is set equal to 8 km, and the reference height $z_r$ is taken to be 12 km. The perturbation fields shown in Figs. 7 and 9 suggest that the direct impact of the imposed heating on the temperature field is very weak above 12 km. As expected, Fig. 13 shows that the impact of the background-state static stability is the most pronounced of the three factors. Therefore, the combination of these three factors, shown in Fig. 13d, closely follows the profile in Fig. 13c. Above 20 km, the rate of increase in $|\phi_z(z)|$ gradually declines. This profile shows qualitative agreement with the simulated temperature profile over the western Pacific (e.g., see day 9 in Fig. 7) where (12) is valid. This finding suggests that the large Kelvin wave temperature perturbation in the LS can be attributed to the strong static stability in the LS region.

If $m$ is allowed to vary as governed by $d_m(x, z, T)/dt = -\alpha U(x, z, T)/\partial z$, with the following three conditions: (i) the inequality represented by (11), where $L$ and $D$ are the horizontal and vertical scales over which $m$ varies; (ii) $O[(1/\kappa)(\partial \phi_\alpha/\partial z)] \ll O[(1/\kappa)(\partial U - N/m_t)/(\partial z)]$; and (iii) $O(\partial U/\partial z) \ll O[(\partial N/m_t)/(\partial z)]$, one arrives at the relation

$$m^2(z) = m^2(z)N(z)/N(z),$$

The factor $[(c - U(z))/[c - U(z)]^{1/2}N(z)/N(z)]^{3/2}$ in (13) becomes $[N(z)/N(z)]^{3/2}$, using (1) and the above relation for $m$ with $z$ being switched to $z$. Thus, the wave energy now depends on $N$ to the power of $5/4$ rather than $3/2$, but still becomes larger with increasing $N$.

Above 25 km, the density factor becomes increasingly important. However, the time scale associated with radiative damping becomes comparable to the time that the Kelvin wave packet takes to reach this level. Because the mean vertical group velocity between 15 and 25 km is about $7.0 \times 10^{-3}$ m s$^{-1}$ (Fig. 12a), it takes approximately 17 days for a Kelvin wave packet to propagate from 15 to 25 km. Given that the radiative damping time scale is about 5 days in the LS (Jin and Hoskins 1995; Taguchi 2003), one expects that wave attenuation resulting from radiative damping will be significant as the wave packet propagates vertically. Because regions higher than 25 km are beyond the scope of this study, these processes are not further considered.

6. Conclusions and discussion

Both regression analyses with global observational datasets and numerical model simulations show that
tropical tropopause undulation is a robust response to convection over the western tropical Pacific. In response to the convective heating, there is warming in the upper troposphere (UT) and cooling in lower stratosphere (LS). With the exception of one region, immediately to the west of the heating, this UT warming and LS cooling occur over most longitudes, and results in a lifting of the tropical tropopause. The model results show that these UT–LS temperature anomalies are closely associated with vertically propagating internal Kelvin waves. As the phase of this internal Kelvin waves progresses downward, the tropopause undulates. The tropopause rises (descends) when cold (warm) Kelvin wave perturbations position themselves immediately above the tropopause.

Based on an analysis of wave action conservation, we suggest that Kelvin waves play a prominent role for the tropical tropopause undulation because temperature perturbations associated with the Kelvin wave attain large values in the vicinity of the tropopause where the static stability is strong. The background zonal wind must also modulate the amplitude of the temperature perturbation, but the amplitude dependency on static stability is much greater.

The model simulations also show that the LS Kelvin wave response is greatest over the western Pacific, regardless of the location of the heating. This result indicates that the LS Kelvin wave attains its largest amplitude over the western Pacific because of wave modulation by the zonal wind of the background state, and not because of the proximity to the heating. Thus, the effect of the heating on the LS Kelvin wave amplitude is realized indirectly through the impact of the heating on the background zonal wind field. This wave modulation in the zonal direction is again consistent with the wave action conservation law, which predicts that, following a zonal ray, the Kelvin wave energy must be proportional to \( \frac{c}{H^2} \). Because the strongest easterlies are found over the western Pacific, the highest wave energy must occur in this region.

There are two nontrivial consequences that arise from the above vertical and zonal modulation of the Kelvin wave: First, a comparison between simulations with the DJF climatological flow and a zonal average of this state indicates that the Kelvin wave modulation in the zonal direction causes a greater amount of zonal mean tropopause displacement. Second, the physical link between the tropical convection, vertically propa-
gating Kelvin waves, and tropopause height has implications for the long-term cooling trend of the tropical cold-point tropopause identified by Zhou et al. (2001). These authors provided evidence that the cooling trend is associated with rising sea surface temperature and increasingly frequent and/or stronger convection. They suggested that the tropical tropopause cooling may occur through the mechanism of Highwood and Hoskins (1998) in which the cold tropopause over the western Pacific is caused by thermal adjustment to the Gill-type circulation. In this picture, stronger and/or more numerous convective activity causes a stronger Gill-type response, and thus increased cooling. The findings in this study indicate that the cooling may not only arise through the mechanism of Highwood and Hoskins (1998), but also through the excitation of vertically propagating Kelvin waves.

More importantly, these results may explain the formation of the cold UT–LS region over the western Pacific, known as the cold trap. As Randel and Wu (2005) pointed out, while the low temperature is also consistent with cooling caused by convective detrainment (Sherwood and Dessler 2001; Kuang and Bretherton 2004), Kelvin waves can also account for the cooling. Tsuda et al. (1994) estimate that the cooling associated with Kelvin waves over the western Pacific may be sufficient to bring the LS air temperature very close to the value necessary for freeze-drying in the cold trap. The wave action analysis presented in this paper indicates that the large Kelvin wave response over the western Pacific is caused by thermal adjustment to the Gill-type cooling may not only arise through the mechanism of Highwood and Hoskins (1998), but also through the excitation of vertically propagating Kelvin waves.

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