The Annual Temperature Cycle of the Tropical Tropopause: A Simple Model Study

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Abstract

A simple radiative-convective model is used to simulate the annual temperature cycle near the tropical tropopause and lower stratosphere (TTL region). Seasonally varying residual vertical velocity and ozone variation are imposed, the latter, derived from 7 years (1998-2004) of southern hemisphere additional ozonesonde (SHADOZ) data. Convection is treated only by eliminating lapse rates greater than 6.5 K/km. An upwelling rate proportional to the extratropical wave driving (mid-latitude EP flux) is sufficient to explain in detail the annual cycles of TTL temperature above 80 hPa and of tropopause pressure, each maximizing in northern summer and minimizing in northern winter. However, temperatures below 80 hPa lag those predicted, indicating either a delay in upwelling or the influence of tropospheric convection. The annual cycle of ozone in the TTL plays an important role in modulating that of temperature: without ozone variations, the simulated temperature amplitude at 70 hPa falls from ~8K to 5K, and the maximum temperature occurs in July, one month earlier than observed. When the seasonal cycle of ozone is included in the calculation, the amplitude and phase of the temperature cycle come into close agreement with observations. These results support the high-latitude wave-driving hypothesis for explaining tropical upwelling, but indicate complicating factors close to the tropopause, and an important role for ozone in modulating temperature changes.
1. Introduction

The tropopause is a basic feature of the atmospheric thermal structure. As the interfacial layer between the troposphere and the stratosphere, it divides the characteristics of both spheres chemically, dynamically and radiatively. The tropopause is linked with many atmospheric features such as the stratospheric circulation, ozone and water vapor distributions, and tropospheric convection. The humidity and thermal structure near the tropical tropopause are important for climate, constituent entry to the stratosphere, and the photochemistry of ozone. The largest variation in the thermal structure of the TTL (tropical tropopause layer) is the long-recognized annual cycle in height and temperature of the tropical tropopause. Analyzing radiosonde data, Reed (1965) found 80 hPa temperatures to be 10 K lower in February than in August. Reid and Gage (1981) also showed that the tropical tropopause is ~1-2 km higher and 5 K colder in Northern hemisphere winter than in summer.

Some authors have tried to explain the annual cycle of temperature near the TTL qualitatively (Reed and Vlcek 1969; Reid and Gage 1981, 1996). Reed and Vlcek(1969) argued that the annual temperature cycle of the lower stratosphere is caused by the annual variation of the adiabatic cooling due to tropical upwelling induced by Hadley circulation penetration into the stratosphere. They assumed that the Hadley circulation is strong in January and weak in July. Reid and Gage (1981) supported this hypothesis, arguing that the annual cycle of tropopause temperature is in response to that of insolation at the surface. They noted that the equivalent potential temperature at the surface matches the potential temperature of tropopause, arguing that the two are linked by maritime convection.
However, Yulaeva et al. (1994) revealed that when the tropical lower stratosphere was the coldest (warmest), the extratropical lower stratosphere was the warmest (coldest). They concluded that the tropical lower stratospheric cooling is not caused by local tropospheric circulations but by the stratospheric circulation. Rosenlof (1995) also found that the upward mass flux in the tropical lower stratosphere is larger during northern hemisphere winter than during northern hemisphere summer, and this annual cycle of mass flux of the lower stratosphere both tropics and midlatitude is controlled by midlatitude momentum forcing. In addition, Reid and Gage (1996) argued that the penetration of overshooting convective turrets reinforced by the Brewer-Dobson circulation caused higher and colder tropopause in the northern hemisphere winter.

The upward propagation and breaking of Rossby and gravity waves transfers zonal angular momentum to the stratosphere, driving the so-called Brewer-Dobson circulation (Haynes et al. 1991). This residual circulation includes downward motion in the midlatitudes, and upward motion in the tropics as compensation. The rate of momentum forcing can be quantified as the divergence of the Eliassen-Palm (EP) flux. Randel et al. (2002a) showed that the EP flux divergence of the northern hemisphere attains its maximum in January and minimum in July, and is roughly sinusoidal. In the southern hemisphere the EP flux divergence maximum occurs in October with a minimum in austral summer. The wintertime EP flux divergence of the northern hemisphere is stronger than that in the southern hemisphere since planetary wave activity, the main source of EP flux, is due to land-sea contrast and orography. When the EP flux divergence of the northern hemisphere extratropics is maximum in January, residual vertical velocity is also maximum.
and the TTL experiences strong adiabatic cooling, and vice versa in the northern hemisphere summer. Thus the annual temperature cycle is antiphase with the EP flux divergence in the northern hemisphere extratropics.

Due to the above studies, the view that the Brewer-Dobson circulation controls tropopause temperature has become dominant. There is a significant problem however in that the upwelling driven by “downward control” occurs immediately equatorward of the applied momentum forcing, or wherever this forcing increases with latitude (Haynes 1991). It is therefore not clear how momentum forcing at high latitudes can remotely influence upwelling over the equator. Indeed, several authors have proposed that tropical upwelling is caused by low latitude momentum sources (Highwood and Hoskins 1998; Boehm and Lee 2003; Norton 2006). On the other hand, a small amount of horizontal mixing may be sufficient to extend high-latitude control into the tropics (Plumb and Eluszkiewicz 1999). In this study we propose to look at the seasonal cycle more carefully than in previous studies, to see if this can discriminate between the hypotheses.

In addition to stratospheric circulation, the radiative effect of chemical compositions can influence the seasonal cycle of temperature near TTL. The ozone and water vapor distributions are closely connected with the thermal structure of the TTL. The seasonal variation of lower stratospheric water vapor is controlled to leading order by the seasonal cycle of temperature at the tropical tropopause (Weinstock et al. 1995). Thuburn and Craig (2002) concluded that the cold point and convection top temperatures are affected by ozone and water vapor due to their radiative effects, but did not address the seasonal cycle.
Tropical upwelling is too small to measure directly. Therefore previous studies have calculated it theoretically from either momentum or thermodynamic balance (Rosenlof 1995; Randel et al. 2002b), or have inferred it from spatio-temporal variations of atmospheric composition (Mote et al. 1996; Andrews et al. 1999).

We use a one-dimensional, radiative convective model in this paper to investigate the seasonal temperature cycle near TTL. Radiative-convective models have been widely used to study the radiative balance of TTL (e.g. Gettelman et al., 2004; Corti et al., 2005) and its sensitivity to atmospheric trace gas and cloud (e.g. Sinha and Shine, 1994; Thuburn and Craig, 2002). Such models capture the fundamental physical processes operating on the largest spatial scales, though they can not resolve phenomena such as equatorial waves. We calculate the residual vertical velocity from thermodynamic balance, and simulate the temperature cycle to get the role of ozone using a one-dimensional, radiative-convective model.

We will first analyze the observed data in section 2, followed by description of the model, and its radiative equilibria. In section 3, we determine the upwelling rate and examine its relation to midlatitude momentum forcing. We simulate, in section 4, the annual temperature cycle with and without the radiative effect of the annual ozone cycle, to ascertain the role of ozone in bringing about the observed behavior. Our results are summarized in section 5.

2. Data analysis and Model description

We used over 1000 temperature and ozone profiles from the Southern Hemisphere
Additional Ozonesonde (SHADOZ) project during the period 1998-2004. The SHADOZ project was established to maintain tropical ozonesonde observations consistently at a few key sites, as opposed to the more sporadic observations available previously (Thompson et al. 2003). Table 1 lists the SHADOZ stations and locations used in this study. Note that only 10 stations within 20° S and 20° N that began collecting data by 1999 are included in this study. Most stations collected data once per week, but some launched only two sondes per month. Balloon-borne ozonesondes fly together with a standard radiosonde for measuring air temperature, pressure, and humidity. As implied by the name, most SHADOZ stations are south of the equator, whereas our model will be intended to represent the deep tropics of both hemispheres equally. This is investigated below.

Figure 1 shows the annual variation of zonal mean temperature (a) and ozone (b) (each layer annual mean is removed and shown at right side axis) calculated from monthly mean of all SHADOZ data. The coldest temperature is in the northern hemisphere winter and the warmest is in summer especially in August as is well known. At 70 hPa, the lowest temperature (in January) is 3.5 K lower than mean of 199.6 K, while the highest (August) is 5 K higher than mean. The amplitude of the annual cycle is 8.5 K. At 100 hPa the peak to peak amplitude is about 4 K which is half of that at 70 hPa. Rapid decreases of the annual amplitude below 100 hPa and above 50 hPa are also shown. This feature can be explained by the radiative restoring timescale: from 100 hPa to 50 hPa this time scale is very long requiring large temperature excursions to balance changes in the energy budget forced by the Brewer-Dobson circulation (Randel et al. 2003).

The ozone concentration and partial pressure have minima near the cold point and
increase sharply with altitude through the lower stratosphere. The ozone mixing ratio is 0.1 ppmv at 100 hPa and 0.5 ppmv at 70 hPa. The annual cycle of ozone resembles that of temperature. The cycle amplitude of ozone mixing ratio is greater at 70 hPa than at 100 hPa, with a maximum in the late Boreal summer and a minimum in winter. However, ozone shows about a one-month time lag compared to the annual temperature cycle.

The comparison of temperature profiles in January (a) and July (b) with NCEP/NCAR reanalysis (Kalnay et al. 1996) data (20°N-S, 1998-2004) is shown in Figure 2. The two data sets agree well except near the cold point in January, even though SHADOZ sites are usually located in southern hemisphere. This is because the difference of the tropical temperature between northern and southern hemisphere is quite small (Figure 2). The significant difference of the temperature near cold point in January is due to lack of the vertical resolution in the NCEP/NCAR reanalysis data. This discrepancy does not appear in July because the cold point happens to be located near the 100 hPa, reanalysis level. We conclude that the SHADOZ network reasonably samples the temperature cycle. Further, tracer concentrations at and above 70 hPa show similar symmetry (e.g., Randel et al. 2001) due presumably to mixing, so SHADOZ ozone should be representative above the TTL. Near the tropopause, meridional ozone sampling biases cannot be ruled out.

The zonal mean energy and momentum budgets are governed by the Transformed Eulerian-mean (TEM) equations (see Haynes 1991). We employ a crude approximation to the energy equation. The flux divergence term in this equation is small under quasi-geostrophic scaling, and Vincent (1968) found the meridional advection term to be small in the lower stratosphere. Ignoring these terms leads simply to:
\[
\frac{\partial \widetilde{\theta}}{\partial t} + \overline{w} \ast \frac{\partial \widetilde{\theta}}{\partial z} = \overline{Q}
\]  

(1)

Where \( \overline{Q} \) is the net diabatic heating (here assumed to be purely radiative), and the other symbols have their usual meanings. In considering this equation, we treat only the “tropical mean,” which is determined by averaging as before over the 10 SHADOZ sites.

To explore radiative heating as simply as possible we employ a one-dimensional, radiative-convective model. This model behaves as a single thermodynamic system under the joint control of radiative and convective processes, with a specified lapse rate determining the tropospheric thermal structure. To calculate radiation we used the Column Radiation Model (CRM), which is a stand-alone version used in version 3 of the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM3) (http://www.cgd.ucar.edu/cms/ccm3/). The CRM treats incoming short wave and outgoing long wave radiation separately. Short wave is resolved by 19 spectral wave bands from 0.2 to 5 \( \mu m \) and 8 bands from 6.5 to 20 \( \mu m \) for long wave. The model was implemented here with 139 layers. These layers have constant pressure thickness of 50 hPa from the surface to 250 hPa, and 2 hPa thickness above this. Since standard radiosondes do not measure water vapor reliably in the TTL, an averaged water vapor profile from Central Equatorial Pacific Experiment (CEPEX) campaign (Kley et al. 1997) is used, and clear sky conditions only are considered. Carbon dioxide is assumed to be well mixed and taken by single mixing ratio of 355 ppmv because the sensitivity to carbon dioxide is very small in range of seasonal cycle (~ 10ppm, Strahan et al. 1998) (Thuburn and Craig, 2002).

Simple convective adjustment to a prescribed lapse rate is used in this study,
following e.g. Manabe and Strickler (1964). The lapse rate at each layer is calculated at
every time step, and if it exceeds 6.5 K/km, it is assumed that two layers exchange heat
until their lapse rate reaches 6.5 K/km. This routine is repeated until the lapse rate of all
layers satisfies this condition. We use a 3 hour time step, but the solar radiative heating rate
is diurnally averaged before it is applied in order to attain a steady solution with no diurnal
cycle. Since the net vertical mass flux between pole to pole must be zero when integrated
along an isobaric surface, the area-averaged vertical velocity should be zero. Because only
the Tropics are treated in this research this constraint obviously does not apply; we compare
our calculated residual vertical velocity to Rosenlof’s result, which was obtained at all
latitudes.

Equilibrium and observed temperature profiles are shown in Figure 3a. The two
equilibrium temperature profiles are calculated with no vertical velocity, and with observed
ozone from January and August as input (Figure 3b). The equilibrium profiles (solid line)
near TTL are far from observed temperature profiles (dash line). They each have a lower
and warmer tropopause than observed, and a significantly warmer lower stratosphere. The
increased ozone in August makes TTL temperatures 2~3 K hotter than they would be if the
ozone were the same as in January. Most of the remaining discrepancy between observed
and equilibrium temperature at upper troposphere can be ascribed to the idealized lapse rate
(6.5 K/km) used here. Near 200 hPa, for example, the average observed lapse rate is almost
7 K/km. The annual cycle of temperature and other features shown in this paper are
practically unaffected by this modification.

The radiative relaxation time scale is about 30 days in the TTL and lower
stratosphere (Fels 1982). For changes on time scales shorter than this, adiabatic cooling variations will tend to be compensated by the local temperature tendency \( \frac{\partial T}{\partial t} \). On annual time scales, however, changes in adiabatic cooling must be balanced largely by those of radiative heating. To facilitate such heating, temperature should be below radiative equilibrium where vertical motion is upward.

3. Residual vertical velocity

Figure 4a shows the annual cycle of the terms in TEM thermodynamic energy equation (1) at 70 hPa. The vertical heat advection (adiabatic heating) and radiative heating have opposite phase and the temperature tendency term is relatively small between two large canceling terms. Figure 4b compares several calculations of the annual variation of vertical velocity \( \bar{w}^* \) at the 70 hPa level. The solid line shows the residual vertical velocity obtained by solving equation (1) for \( \bar{w}^* \) and inserting observed temperature and heating rates calculated from them. This estimate peaks in early northern hemisphere winter and reaches a minimum in summer. The dash-dot line shows the result of Rosenlof (1995), which was calculated by the same method but using different data (in particular, satellite retrievals of ozone). These two calculations are in good agreement with respect to peak-to-peak amplitude, though our calculation shows an earlier peak of \( \bar{w}^* \) in NH winter and fall-off in NH spring.

If we recalculate \( \bar{w}^* \) without taking into account the seasonal variation of ozone (using January ozone for the entire year), we obtain the result shown as dashed line in the
figure. In this case the seasonal variation of diagnosed $\bar{w}^*$ is much greater. This is because we no longer account for the higher ozone in NH summer, which increased the net heating rate and therefore the diagnosed ascent making it closer to the strong winter ascent. Quantitatively, the presence of an annual cycle in ozone increases the diagnosed residual vertical velocity by 10 m/day in July compared to January.

Unfortunately we have no direct observations of $\bar{w}^*$ averaged over the tropics to which these estimates may be compared. However, we may examine the hypothesis that the variations are driven by midlatitude wave-driving, by comparing the seasonal variations of those two quantities. As an index of wave-driving, we examine the EP fluxes calculated in the lowermost stratosphere. Using NCEP/NCAR reanalysis data, we calculated the monthly mean $\bar{\nu'T'}$ for the period from 1998 to 2004. The zonally averaged eddy heat flux $\bar{\nu'T'}$ should be a good proxy for EP flux, since EP flux is proportional to eddy heat flux under quasi-geostropic conditions (Andrews et al. 1987, Randel et al. 2002). Previous work has related changes in EP flux to variations in ozone, in a manner consistent with dynamical control of the Brewer-Dobson circulation by the EP flux on intraseasonal and longer time scales (Newman and Nash, 2000; Randel et al., 2002)

Figure 5 shows the monthly mean 100 hPa $\bar{\nu'T'}$ at 40-70° latitude in the northern hemisphere (a) and southern hemisphere (b) for 1998-2004. In the northern hemisphere, the maximum occurs in January and the minimum in July. In the southern hemisphere, the maximum is in October and the minimum is in January, though the amplitude (10 K·m/s) is lower than that in the northern hemisphere (15 K·m/s). One expects the Tropics to be affected equally by driving from both hemispheres. Accordingly figure 5 (c) and (d)
compares the sum over both hemispheres (solid lines) with the previous estimated $\overline{w}^*$ (dashed lines, the same as the solid lines in figure 4b) at 70 hPa (c) and 100 hPa (d). The scaling and zero-offset of the $\overline{w}^*$ curve (right axis) were arbitrarily chosen to yield similar minimum and maximum position of the two curves.

The match between the two curves at 70 hPa is very encouraging, indicating that the high-latitude EP flux does indeed appear to be a sufficient explanation for the seasonal variations of tropical ascent. However, at 100 hPa, the residual vertical velocity is delayed compared to the EP flux. Furthermore, comparing the right axes of the two panels, one sees that the annual range of upward mass flux (proportional to $\overline{w}^* p$) is similar at both levels, but $\overline{w}^*$ has a constant offset at 100 hPa that is comparable to the annual range. In other words, there seems to be another agent driving upward motion throughout the year in addition to that attributable to midlatitude wave driving.

Table 2 shows the correlation between heat fluxes of several latitude ranges and pressure levels and the residual vertical velocity of 70 hPa in tropics. The correlation coefficient increases when the latitude range is higher, and in a given latitude range, results are similar from 150 hPa to 50 hPa. At latitudes from 20-30°, the correlation coefficient is even negative when the pressure level is lower than 50 hPa. This indicates that upwelling in the tropics is firmly connected to extratropical forcing.

Figure 6 shows the correlation coefficient (a) and lag (b) between the residual vertical velocities from 100 hPa to 50 hPa and the monthly mean $\overline{\nu' T'}$. For the vertical velocity time series at a given pressure, we first computed the correlation coefficient against the heat flux series from each combination of the four pressure levels and two
midlatitude (40-70° and 50-80°) ranges, then averaged the eight coefficients. The 1-sigma error bars are shown. The residual vertical velocity is well correlated with the EP flux \( r > 0.6 \) at all pressure levels. Especially, the correlation coefficient is over 0.8 at the range of 80-60 hPa. The lag correlation is somewhat interesting. Small lag (less than 10 day) was found above 80 hPa, however, an interesting lag does appear below 80 hPa, reaching 34 day at 100 hPa, indicating that the second agent driving ascent probably originates in the troposphere.

We thus have an additional dynamical agent localized near the tropopause. One possibility is momentum transport by Rossby waves driven by zonally asymmetric tropical heating, a mechanism similar to mid-latitude wave driving but originating in the Tropics (e.g., Boehm and Lee 2003). Recently Norton (2006) proposed a similar mechanism. A different possibility is that ascent diagnosed based on (1) is too large near the tropopause because an energy sink due to overshooting deep convection (Sherwood 2000; Kuang and Bretherton 2004).

Our calculation ignores the effects of clouds, which affect net heating rates and cloud, in principle, provide another energy sink (or source). Optically thick clouds cause cooling in the TTL by reducing upwelling thermal radiation (Dessler et al. 1996). However, this small effect is about the same above and below 80 hPa so it cannot contribute to the behavior change near the tropopause noted here. Cooling at the tops of optically thick clouds would be more localized, but makes a negligible contribution due to low coverage near the tropopause (Sherwood 2000).

Thin cirrus clouds have greater coverage and may have significant effects.
Hartmann et al. (2001) showed that these decrease the radiative heating if they are located above an anvil, but increase it (by a much greater amount) if they are not. Observations indicate that cirrus cover ~30% of the Tropics and that those present are located over optically thick, deep clouds less than half the time (McFarquhar et al 2000, Wang and Dessler 2005). Thus the heating effect will dominate. McFarquhar et al. (2000) estimate that the average cirrus is 0.5 km thick and heats internally at 1-2 K/day. If this is averaged over the Tropics and over a ~3 km layer in which thin cirrus occur, the resulting contribution is a net heating of 0.05-0.1 K/day. This drops slightly if offsetting cooling by dual-layer clouds is considered, but could also be bigger if optical depths are underestimated. Evidently cirrus effects are modest compared to the clear-sky heating rate of ~0.5K, and if anything, increase the discrepancy in behavior between the TTL and lower stratosphere rather than account for it.

The two remaining possibilities, convective cooling or additional ascent, are not mutually exclusive and might both contribute to the observed effect. In any case we can infer that the residual vertical velocity calculated by the thermodynamic equation is sufficient over 80 hPa, but below 80 hPa, there may be other terms in the energy budget besides radiation.

4. Contribution of ozone to annual temperature cycle

In this section, we simulate the temperature annual cycle with the residual vertical velocity and one-dimensional, radiative-convective model. As mentioned before, the residual vertical velocity is calculated with the observed temperature and heating rate
calculated by the same model. Thus we will trivially recover the seasonal variation in temperatures used in (1) to diagnose in the first place, including those at 100 hPa where there are questions about closure of the heat budget. However, this methodology allows us to investigate the sensitivity of temperature variations to the treatment of ozone or water vapor. The conclusions will be insensitive to how the imbalance below 80 hPa is ultimately resolved, provided that the missing process (convection, tropical wave breaking, or something else) has a reasonably steady impact throughout the year when averaged over the Tropics. Figure 7 shows the results at 70 hPa (a) and 100 hPa (b). The solid line indicates the monthly mean temperature annual cycle with ozone changes, and the dashed line is the cycle without ozone changes. The dash-dot line shows the observed monthly mean temperature annual cycle. No water vapor changes are included.

As in Section 2, there are small discrepancies in average temperature that can easily be attributed to imperfections in our assumed condition of convective neutrality (6.5 K/km lapse rate). With varying ozone, as expected, the seasonal variation of the original temperature data is recovered to very good approximation.

The dashed line in figure 7 shows the result with ozone and water vapor fixed over the entire year. The temperature peaks in JJA, one month earlier than in the observation and the variable-ozone simulation. The maximum temperature is 202.32 K at 70 hPa (2.6 K lower than in the control calculation) and 197.26 K at 100 hPa (1 K lower than the control). Thus the radiative effect of annual ozone variations increases the annual temperature range in the lower stratosphere by about 2K or 35%. In addition to temperature structure, the annual cycle of cold-point pressure is also affected by that of ozone. The difference of the
cold-point pressure in northern hemisphere winter and summer is about 14 hPa (86 hPa in January and 100 hPa in August) in observations and in our control calculation. However, this amplitude is reduced to 12 hPa in the fixed ozone case (figure is not shown). Figure 8 shows the time lag calculated from two simulated annual temperature cycle (with, and without the ozone change) compared to the residual vertical velocity. The temperature cycle lags that of upwelling by from 10 days (100 hPa) to 35 days (70 hPa ~ 60 hPa) due to the radiative restoring timescale as mentioned in section 2. In addition to this time delay, the annual ozone cycle lags that of temperature by additional time from 5 days (100 hPa) up to 25 days (70 hPa). Consequently, the observed temperature lags the residual vertical velocity or EP flux by up to two months. The impact of variable ozone is qualitatively similar at the 100 hPa and 70 hPa levels.

The annual variation of radiative heating due to ozone at 100 hPa, up to 0.03 K/day or roughly 10 % of net clear sky heating rate, results in residual vertical velocity increases of 3.5 m/day. The uncertainty in ozone impacts is thus fairly small, of similar order to the anticipated contribution from thin cirrus discussed earlier. We also examined the importance of seasonal changes in water vapor, by imposing a large seasonal variation. This experiment (not shown) produced negligible impacts compared to those of ozone, indicating that water vapor is neither a critical factor in the seasonal temperature cycle nor a significant source of uncertainty in mean heating rate. This is consistent with the conclusions of Thuburn and Craig (2002).

Thus, the combined influences of residual vertical velocity and ozone are sufficient for understanding the annual temperature cycle in the lower stratosphere. However, the
behavior below 80 hPa requires further attention, since the residual vertical velocity must include other effects as discussed in Section 3. Despite this limitation, it is likely that the ozone variation amplifies the seasonal temperature variation, and delays it by about a month, throughout the near-tropopause and lower stratosphere region.

5. Conclusion

We have performed both inverse calculations (estimating ascent from the simplified energy equation (1)) and forward ones (simulating temperature variations based on inferred ascent variations) to relate tropical-mean ascent, temperature, tropopause pressure, and ozone concentrations over a mean annual cycle. The inverse calculations yield results for the annual cycle of vertical velocity ($\vec{w}$) similar to those previously reported (Rosenlof 1995), but we find that they depend significantly on the specification of the seasonal cycle of ozone, which we obtained here from seven years of tropical ozone sondes at 10 sites. The new $\vec{w}$ variation compares very favorably with dynamical calculations of midlatitude wave driving, indicating that the annual cycle of lower-stratospheric temperature (above 80 hPa) is fully explained by this forcing to within observational uncertainties. This does not appear to be the case below 80 hPa, where ascent is stronger than expected throughout the year (as found previously by Rosenlof(1995) and others) and lags the forcing by almost one month. Thus, we infer that the residual vertical velocity calculated with thermodynamic equation is distorted near the tropopause by other more local influences. These probably include cooling due to deep convective overshooting and/or local ascent possibly associated with wave-mean flow interactions driven by
asymmetries in tropospheric heating.

Forward calculations of the annual cycle of temperature indicate that there is a one-month time lag between the annual temperature cycle that would result from a fixed ozone condition and the cycle of residual vertical velocity at 70 hPa. This lag does not agree with the nearly two-month lag inferred from the observed temperature and wave-driving at 70 hPa. It is the annual cycle of ozone in the TTL that produces the additional time lag of up to one-month, as well as a 3-4 K increase in the seasonal temperature range in the lower stratosphere. We have to note that we use only 10 stations of SHADOZ project which have only two northern hemisphere stations. This means that our average temperature and ozone may not represent whole tropics, even though the temperature bias does not appear to be important based on reanalysis data, nor are large interhemispheric ozone differences indicated in satellite retrievals. We conclude that ozone is an important factor, in addition to Brewer-Dobson circulation, in explaining the annual temperature cycle of the TTL.

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Table and Figure

Table 1. SHADOZ stations and location

Table 2 Correlations between heat fluxes of several latitude ranges (both northern and southern hemisphere) and pressure levels and the residual vertical velocity at 70 hPa in tropics

Figure 1. Annual Cycle of SHADOZ (Annual average of each pressure level is removed in contours, and showed at the right side axis.) (a) Temperature. Contour interval is 0.5K (b) Ozone. Contour interval is 0.02ppmv. Solid lines of both figures indicate positive and dash is negative.

Figure 2. Vertical temperature profiles of averaged SHADOZ (thick line) and NCEP/NCAR reanalysis over 20°N – 20°S from 1998 to 2004 (thin line). Northern (0°-20°N, asterisk), and southern hemisphere (0°-20°S, diamond) temperature profiles are shown.

Figure 3. (a)Equilibrium temperature profile in which January and August mean ozone are input (line) and respective SHADOZ temperature profile (dash line). (b) January and August ozone profiles. (each left side among pair lines is January profile.)

Figure 4. (a) all terms in the thermodynamic energy equation (1) at 70 hPa. Solid line is
the temperature tendency, dash line is the radiative heating, and dash-dot is the vertical temperature advection (adiabatic heating). (b) Annual cycle of residual vertical velocity diagnosed from (1) at 70 hPa using SHADOZ temperature and ozone, with ozone seasonally varying (solid) and fixed at its January profile (dashed). Dash-dot is the result of Rosenlof (1995).

Figure 5. (a) Time series of monthly averaged 100 hPa $\vec{v}T$ (Km/s) over the northern hemisphere (40-70N) for 1998-2004. Individual lines refer to individual years. (b) The same $\vec{v}T$ over the southern hemisphere (40-70S). The sign is reversed. (c) Added $\vec{v}T$ of both hemisphere (thick solid line), and the residual vertical velocity calculated in section 3 (dashed line) at 70 hPa (d) The same as (c) except the residual vertical velocity at 100 hPa.

Figure 6. The correlation coefficient (a) and lag (b) between the residual vertical velocity calculated by thermodynamic equation and $\vec{v}T$. The 1-sigma error is also shown.

Figure 7. Annual temperature cycle calculated with seasonally varying ozone (solid line), with fixed (January) ozone (dashed). Observed cycle is also shown (dash-dot). (a) 70 hPa (b) 100 hPa.

Figure 8. The lag between the residual vertical velocity and temperature with variable ozone (solid line) and with fixed ozone (dashed).
Table 1. SHADOZ stations and location

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</table>
Table 2 Correlations between heat fluxes of several latitude ranges (both northern and southern hemisphere) and pressure levels and the residual vertical velocity at 70 hPa in tropics

<table>
<thead>
<tr>
<th>Pressure</th>
<th>latitude</th>
<th>20°-30°</th>
<th>30°-60°</th>
<th>40°-70°</th>
<th>50°-80°</th>
</tr>
</thead>
<tbody>
<tr>
<td>150 hPa</td>
<td>-0.533</td>
<td>0.736</td>
<td>0.906</td>
<td>0.952</td>
<td></td>
</tr>
<tr>
<td>100 hPa</td>
<td>-0.844</td>
<td>0.817</td>
<td>0.933</td>
<td>0.942</td>
<td></td>
</tr>
<tr>
<td>70 hPa</td>
<td>-0.598</td>
<td>0.874</td>
<td>0.928</td>
<td>0.939</td>
<td></td>
</tr>
<tr>
<td>50 hPa</td>
<td>0.284</td>
<td>0.878</td>
<td>0.911</td>
<td>0.924</td>
<td></td>
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</tbody>
</table>
Figure 1. Annual Cycle of SHADOZ (Annual average of each pressure level is removed in contours, and showed at the right side axis.) (a) Temperature. Contour interval is 0.5K (b) Ozone. Contour interval is 0.02ppmv. Solid lines of both figures indicate positive and dash is negative.
Figure 2. Vertical temperature profiles of averaged SHADOZ (thick line) and NCEP/NCAR reanalysis over 20°N – 20°S from 1998 to 2004 (thin line). Northern (0°-20°N, asterisk), and southern hemisphere (0°-20°S, diamond) temperature profiles are shown.
Figure 3. (a) Equilibrium temperature profile in which January and August mean ozone are input (line) and respective SHADOZ temperature profile (dash line). (b) January and August ozone profiles. (each left side among pair lines is January profile.)
Figure 4. (a) all terms in the thermodynamic energy equation (1) at 70 hPa. Solid line is the temperature tendency, dash line is the radiative heating, and dash-dot is the vertical temperature advection (adiabatic heating). (b) Annual cycle of residual vertical velocity diagnosed from (1) at 70 hPa using SHADOZ temperature and ozone, with ozone seasonally varying (solid) and fixed at its January profile (dashed). Dash-dot is the result of Rosenlof (1995).
Figure 5. (a) Time series of monthly averaged 100 hPa $\sqrt{\mathbf{T}'^2}$ (Km/s) over the northern hemisphere (40-70N) for 1998-2004. Individual lines refer to individual years. (b) The same $\sqrt{\mathbf{T}'^2}$ over the southern hemisphere (40-70S). The sign is reversed. (c) Added $\sqrt{\mathbf{T}'^2}$ of both hemisphere (thick solid line), and the residual vertical velocity calculated in section 3 (dashed line) at 70 hPa (d) The same as (c) except the residual vertical velocity at 100 hPa
Figure 6. The correlation coefficient (a) and lag (b) between the residual vertical velocity calculated by thermodynamic equation and $\sqrt{\nu T}$. The 1-sigma error is also shown.
Figure 7. Annual temperature cycle calculated with seasonally varying ozone (solid line), with fixed (January) ozone (dashed). Observed cycle is also shown (dash-dot). (a) 70 hPa (b) 100 hPa.
Figure 8. The lag between the residual vertical velocity and temperature with variable ozone (solid line) and with fixed ozone (dashed).