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Ozone Transport-Radiation Feedbacks in the Tropical Tropopause Layer

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Key Points:
- Interactive ozone chemistry is applied in a single-column radiative-convective equilibrium model to investigate transport processes
- Ozone transport by Brewer-Dobson circulation upwelling is of first-order importance for temperature structure of tropical tropopause layer
- Transport-radiation feedback: increased (decreased) upwelling reduces (increases) ozone −20% of the total cooling (warming) from upwelling

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Abstract The tropical tropopause layer (TTL) temperature balance is of considerable interest for its control over the amount of water entering the stratosphere. The upwelling branch of the Brewer-Dobson circulation (BDC) directly affects these temperatures through adiabatic cooling. BDC upwelling also indirectly affects TTL temperatures through the influence of ozone transport on radiative heating. We investigate this latter feedback using a single-column radiative-convective equilibrium model coupled with a model of simplified stratospheric ozone chemistry and vertical transport. We find that BDC ozone transport is of first-order importance for TTL temperatures. Additionally, we estimate the effect of ozone transport on cold point tropopause temperature responses to changes in upwelling. We find that the feedback is responsible for approximately 20% of the response to perturbations on time scales longer than about half a year but that this contribution can be neglected for time scales shorter than about a week.

1. Introduction

The Brewer-Dobson circulation (BDC) and tropical tropopause layer (TTL) have considerable effects on the stratosphere and interact through subtle mechanisms, making them topics of particular interest. The BDC (see Butchart, 2014, for a review) is the primary circulation in the stratosphere and therefore plays a critical role in the distribution of stratospheric chemical species, such as ozone (Dobson et al., 1929, 1956) and water (Brewer, 1949; Mote et al., 1996). For example, the BDC is the main process that brings tropospheric water into the stratosphere, carrying it through the TTL. The TTL, meanwhile, is a troposphere-stratosphere transition region, which extends from about 14 to 18.5 km (150-70 hPa) and is horizontally bounded by the subtropical jets (see Fueglistaler et al., 2009, for a review). As air is brought through the tropical tropopause (17 km, 90 hPa) by the BDC, the humidity is controlled by the cold point tropopause temperature via freeze drying (Brewer, 1949). Thereby, the BDC and TTL, working in concert, set the entry value of stratospheric water concentrations, to first order. In turn, stratospheric water concentrations can have significant effects on both climate sensitivity (Solomon et al., 2010), TTL temperatures (Birner & Charlesworth, 2017; hereafter *BC17*), and the BDC (Maycock et al., 2013). The tropical cold point tropopause temperature, although only a small part of the atmosphere, therefore has outsized effects on atmospheric properties and processes beyond both the tropics and the stratosphere.

Unfortunately, climate models show a large spread in tropical cold point tropopause temperatures, on the order of 10 K, which does not have an obvious single cause (Gettleman et al., 2010). This range of model tropical cold point temperatures introduces considerable uncertainty into their representations of stratospheric climate and could be caused by even small differences in model temperature balances in the TTL, due to long radiative time scales in the region (Hitchcock et al., 2010). One process with a strong impact on TTL temperatures is BDC upwelling, which has been connected to temperature within the TTL in terms of both the mean state (e.g., BC17; Thuburn & Craig, 2002) and seasonal variability of the cold point tropopause (e.g., Fueglistaler et al., 2011; Ming et al., 2017). The simplest mechanism for these effects is the direct effect of upwelling on temperature through adiabatic cooling. However, BDC upwelling also exerts an indirect effect on temperature through shaping the TTL ozone profile by vertical ozone transport, which affects local radiative heating rates (Abalos et al., 2013; Avallone & Prather, 1996; Solmon et al., 1985). We refer to this latter effect, separate from adiabatic cooling, as the *ozone transport-radition feedback*. This effect has also

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been previously connected to stratosphere-wide processes with attendant feedbacks on surface climate sensitivity (Nowack et al., 2014). Understanding this process is valuable for understanding the balance that sets temperatures in the TTL.

Single-column radiative-convective equilibrium (RCE) models have proved instrumental in the study of this topic, as the strength of processes can be precisely and individually controlled within these idealized models. Earlier works have shown the importance of convection and local concentrations of radiatively relevant chemical species (Manabe & Strickler, 1964; McElroy et al., 1992; Thuburn & Craig, 2002) for the TTL temperature structure. Recently, the first-order importance of adiabatic cooling from BDC upwelling on the tropical cold point tropopause temperature has also been demonstrated (BC17; Dacie et al., 2019). All of these previous works have performed ad hoc adjustments in prescribed ozone profiles, noting the capacity of ozone alterations to change the lower stratospheric temperature profile. Similarly, BC17 showed that differences in chemistry-climate model ozone profiles could, in principle, explain the large spread in model tropical cold point temperatures. However, no previous single-column RCE study has grounded these adjustments in any ozone-affecting process, let alone studied the ozone transport-radition feedback. Given the noted importance of ozone in TTL temperatures, and the effects of these temperatures on stratospheric climate, the study of this feedback is of considerable interest.

In this study we investigate this feedback by inclusion of a single-column model of simplified stratospheric ozone chemistry and transport into the single-column RCE model of BC17. We find that ozone transport due to the BDC is of first-order importance for the observed TTL temperature structure. We also quantify the contribution of the ozone transport-radiation feedback to the cold point tropopause temperature response to changes in upwelling. We find that this contribution is sensitive to the time scale of upwelling changes. The contribution maximizes with 20% on time scales of 1 year and longer but reduces as the upwelling change time scale is decreased, becoming negligible on weekly time scales. Particularly strong sensitivity of the contribution is found near the TTL radiative relaxation time scale.

2. Model and Methods

This work extends the single-column RCE model of BC17 to solve for two coupled evolution equations of temperature and ozone. These equations are applied simultaneously across the model domain on every time step. The model has 194 layers with high resolution in the TTL (22 layers between 14 and 18.5 km) and a model top height of about 55 km. Simulations are all performed for 10° S latitude and 15 January to maintain consistency with the SHADOZ (Witte et al., 2017) January-mean ozone reference profile and the work of BC17. Also, all simulations include a diurnal cycle in solar zenith angle and use an integration timestep of 1.5 hr.

2.1. The Temperature Balance

Model temperature tendencies are computed through radiation, convection, and the effects of BDC upwelling on temperature (as this component is identical to BC17; see that reference for details). Radiation for heating rates is computed using the Rapid Radiative Transfer Model for GCM Applications, a radiative transfer model that is commonly used as the basis for radiative transfer schemes in modern climate models (Mlawer et al., 1997). All radiative heating rates are calculated for clear-sky conditions. The water vapor profile is calculated using a prescribed relative humidity of 50% and a minimum water mixing ratio (4 ppmv). Convective adjustment is applied with a fixed 6.5 K/km critical convective lapse rate, as in many previous RCE works (BC17; Fu et al., 2018; Manabe & Strickler, 1964; Thuburn & Craig, 2002), and with a fixed 300 K surface temperature. Upwelling is simulated with a prescribed velocity in the layers above the convection top, for which a control value of 0.5 mm/s is assumed. This control value is approximately the upwelling velocity near the tropical tropopause found by Abalos et al. (2015).

This is a simplification of the expected tropical BDC upwelling profile, which strengthens below the tropical tropopause and—to a lesser extent—weakens above it (Fu et al., 2007). This approximation, however, allows for a simple and consistent way to change upwelling near the cold point tropopause.

The model temperature equation is then

$$\partial_t T = Q - \bar{w}^*_T (g/C_p + \partial_z T), \tag{1}$$

where Q is the diabatic heating rate and $\bar{w}^*_T$ is the upwelling velocity applied to temperature calculations. We refer to the second term as the *adiabatic cooling* due to BDC upwelling.

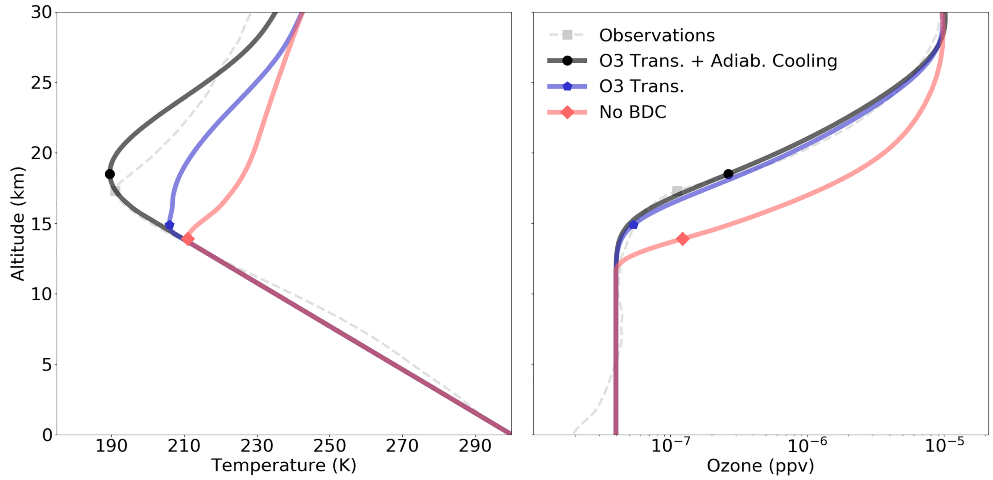

Figure 1. Profiles of temperature and ozone from model results of simulations including both upwelling effects ("O3 Trans. + Adiab. Cooling"), only ozone transport ("O3 Trans."), and no upwelling effects at all ("No BDC"). Observations shown for reference (SHADOZ January climatology, same as in BC17). When included, upwelling is simulated with a value of 0.5 mm/s. Markers indicate the position of the cold point tropopause. Note that tropospheric temperature and ozone are equal for all simulations due to our simple treatment of prescribed SST and tropospheric lapse rate, and prescribed surface ozone mixing ratio (see section 2).

2.2. The Ozone Balance

TTL ozone is mainly shaped by two processes: (1) photolytic ozone production and (2) vertical transport from upwelling (Avallone & Prather, 1996; Abalos et al., 2013; Solmon et al., 1985). Modeling TTL ozone is a primary purpose of our model, so these processes were both accounted for. As the photolytic ozone production rate at one altitude depends strongly on the total ozone column above it, our model must also have some representation of middle and upper stratospheric ozone processes. Above about 25 km in the tropics, chemical ozone loss exerts a first-order control on ozone concentrations (Solmon et al., 1985) and has four main pathways: odd-hydrogen (HOx), chlorine (ClOx), odd-nitrogen (NOx), and pure oxygen (Chapman) losses. However, we have found that the details of this representation are not crucial to obtain realistic TTL photolytic production rates. Therefore, in the interest of reducing model complexity, we have only accounted for the NOx and Chapman loss pathways, which are the most important and simplest pathways, respectively.

Under the conditions that are meant to represent the observed atmosphere (profile "O3 Trans. + Adiab. Cooling" of Figure 1), the model produces an ozone column of 190 DU at 18 km. This is approximately 10% smaller than the reference profile (see "Observations" in the same figure) value of 208 DU.

The reactions accounted for in the photochemical calculation are as follows: $\mathrm{J_{O_2}}: \mathrm{O_2} \rightarrow 2\mathrm{O}$, $\mathrm{J_{O_3}}: \mathrm{O_3} \rightarrow \mathrm{O_2}+\mathrm{O}$, $\mathrm{k_2}: \mathrm{O_2}+\mathrm{O}+\mathrm{M} \rightarrow \mathrm{O_3}+\mathrm{M}$, $\mathrm{k_3}: \mathrm{O_3}+\mathrm{O} \rightarrow 2\mathrm{O_2}$, $\mathrm{k_4}: \mathrm{NO}+\mathrm{O_3} \rightarrow \mathrm{NO_2}+\mathrm{O_2}$, $\mathrm{k_5}: \mathrm{NO_2}+\mathrm{O} \rightarrow \mathrm{NO}+\mathrm{O_2}$.

The first four reactions are part of the Chapman odd-oxygen system. The latter two reactions form the catalytic cycle of NOx odd-oxygen destruction, which have been simplified using a rate limiting step approximation (as in Stolarski & Douglass, 1985).

With this in mind, the ozone equation for this model is

$$\frac{\mathrm{d}\chi_{\mathrm{O_3}}}{\mathrm{d}t} = 2J_{\mathrm{O_2}}\chi_{\mathrm{O_2}} - \frac{2J_{\mathrm{O_3}}k_3}{k_2\chi_{\mathrm{O_2}}[\mathrm{M}]}\chi_{\mathrm{O_3}}^2 - \frac{2J_{\mathrm{O_3}}k_5\chi_{\mathrm{NO_2}}}{k_2\chi_{\mathrm{O_2}}[\mathrm{M}]}\chi_{\mathrm{O_3}} - \bar{w}_{\chi}^{*}\partial_z\chi_{\mathrm{O_3}}. \tag{2}$$

As for temperature, the value of $\bar{w}_{\chi}^{*}$ is prescribed but is applied in the layers at and below the convection top instead of only the layers above the convection top. This represents a slight inconsistency with the temperature balance but is necessary to provide some ozone transport in the troposphere. An alternative choice is to set ozone in layers at and below the convection top to a tropospheric background value, but this can produce rapid changes in the ozone profile as the convection top height can change over the course of a day. Photolysis rates ($J_{\mathrm{O_2}}, J_{\mathrm{O_3}}$) are calculated on every time step. For this, we implemented a radiative transfer calculation with a plane-parallel approximation using band-averaged calculations of the transmission function accounting for ozone and diatomic oxygen. Cross section, quantum yield, and reaction rate

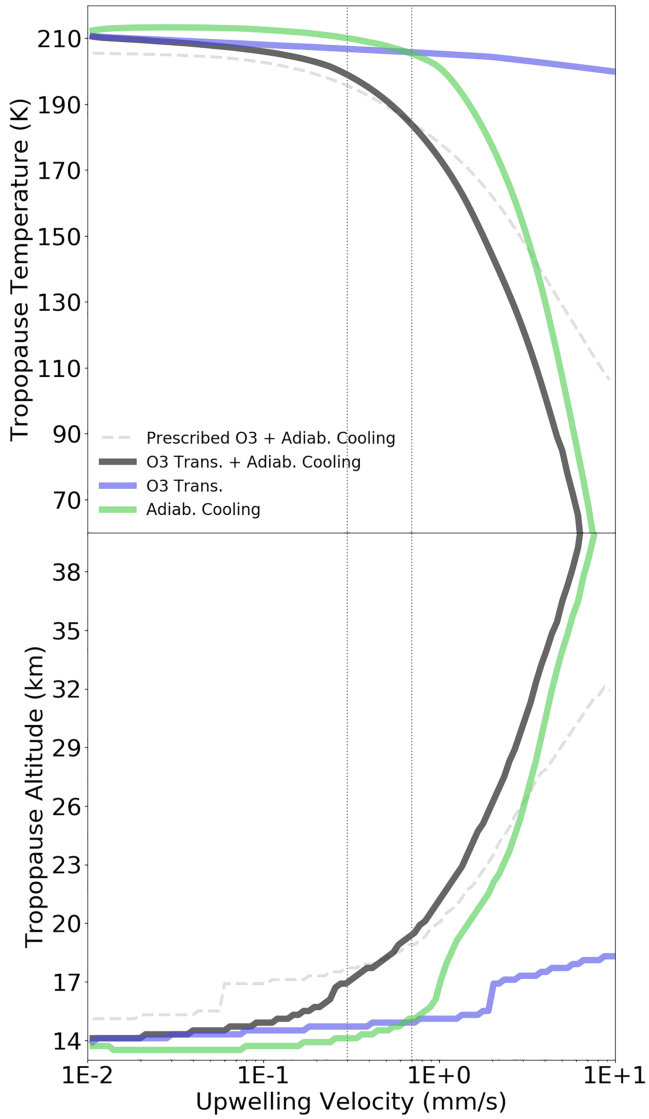

Figure 2. Equilibrium cold point tropopause temperature and altitude versus upwelling velocity. Shown are results from model runs with only adiabatic cooling ("Adiab. Cooling"), only ozone transport ("O3 Trans."), both effects ("O3 Trans. + Adiab. Cooling"), and with ozone prescribed from SHADOZ observations and with adiabatic cooling ("Prescribed O3 + Adiab. Cooling"). Thin dotted lines denote the approximate range of observed upwelling velocity.

($k_2, k_3,$ and k_5) data follow the recommendations of Burkholder et al. (2015), including Yoshino et al. (1988), with an additional parameterization for cross sections within the Schumann-Runge bands (Minschwaner et al., 1993). The data for NO_2 mixing ratios are taken from the ACE-FTS satellite data set (Strong et al., 2008). The surface layer mixing ratio of ozone is fixed at 40 ppbv.

In simulations where ozone transport is absent, obtaining photochemical equilibrium ozone mixing ratios would require excessive integration time (on the order of thousands of years, at least). To reach equilibrium ozone concentrations quickly, a quasi-analytical method is applied. The method requires solving the ozone balance equation (excluding transport) for the equilibrium ozone concentration using the daily-average coefficients of χ. This adjustment is then performed daily for 500 days, after which the model is integrated normally (to ensure radiative equilibrium), achieving the same result as an extremely long integration would. To be clear, this is only performed when the BDC upwelling velocity for ozone transport is zero.

3. Can the TTL Temperature Structure Exist Without BDC Ozone Transport?

Figure 1 shows that our model is capable of qualitatively reproducing the observed temperature and ozone structures of the TTL when the full BDC effects, that is, for both temperature and ozone, are included (black lines, circular marker). In this case, the cold point is well separated from the top of convection (where the temperature structure starts to depart from the prescribed lapse rate), and its temperature and altitude are very close to those of the observed cold point. If the model only includes ozone transport (i.e., no adiabatic cooling, blue lines in Figure 1), the model lower stratosphere is much warmer with a quasi-isothermal layer stretching almost 4 km above its cold point. In this case the cold point is approximately 15 K warmer and more than 2 km lower than observations (nearly colocated with the top of convection). This temperature profile is structurally unlike observations and does not resemble that of the observed TTL. Note that the TTL shows slightly increased ozone mixing ratios compared to when upwelling-induced adiabatic cooling is included (blue vs. black lines). This is due to reduced ozone concentrations above 30 km, which are caused by increased temperatures at these levels and the inverse proportionality of ozone loss rates to temperature (Craig & Ohring, 1958). Reduced upper-level ozone causes increased penetration of shortwave radiation into the lower stratosphere, hence increased production rates creating increased ozone concentrations.

When upwelling is neglected altogether (i.e., no adiabatic cooling and no ozone transport), the ozone profile structure is also unlike observations (red lines in Figure 1). In this case photochemical equilibrium creates lower stratospheric ozone concentrations that exceed observed values by an order of magnitude. These highly elevated ozone concentrations in turn cause increased shortwave absorption and temperatures that exceed observed values by up to 30 K. Compared to the case with only ozone transport (blue lines), the cold point is even warmer and located at a lower altitude, marking a strong temperature inversion. The simulated temperature profile without BDC effects does not resemble that of the observed TTL.

We have also obtained a temperature profile when BDC-induced adiabatic cooling is included but ozone transport is excluded, which is overall similar to the case with only ozone transport (not shown). This indicates that both direct and indirect BDC effects (due to adiabatic cooling and ozone transport-radiative

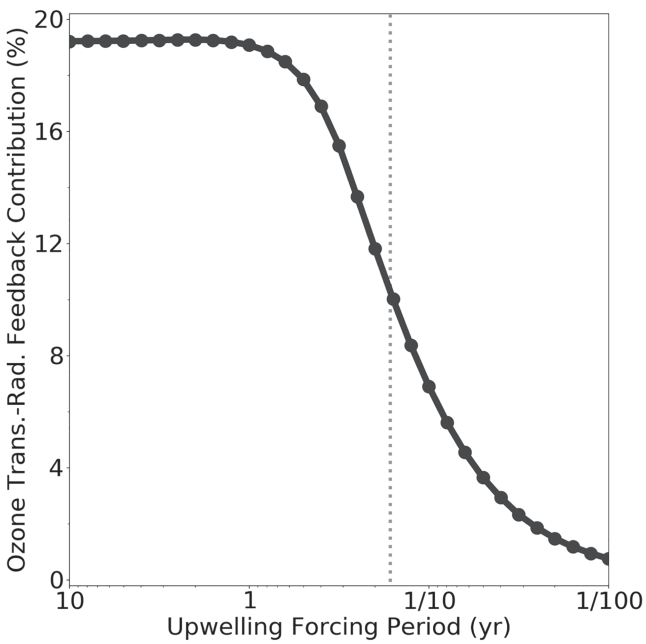

Figure 3. Percentage contribution of the ozone transport-radiation feedback to the cold point temperature response to an oscillating upwelling forcing over a range of upwelling forcing periods. Upwelling forcing amplitude is 0.25 mm/s—half the mean upwelling velocity of 0.5 mm/s. Thin dotted line indicates the approximate TTL radiative relaxation time scale of 60 days.

feedback, respectively) are of roughly equal importance to the temperature structure of the TTL. However, the relative importance of these two effects is a function of upwelling strength, as shown in Figure 2.

Including both effects ("O3 Trans. + Adiab. Cooling") produces colder and higher cold points at all upwelling velocities compared to when only one effect is included ("O3 Trans." or "Adiab. Cooling"). Two regimes of the impact of upwelling on the cold point emerge from these results—the slow upwelling regime (<0.1 mm/s) and the fast upwelling regime (>1.0 mm/s). In the slow upwelling regime, adiabatic cooling seems to be less important than ozone transport, as inclusion of only the former does not show much tropopause temperature sensitivity to upwelling while inclusion of the latter almost matches the tropopause temperature sensitivity when both effects are included. Furthermore, the model run using a prescribed ozone profile (which implicitly accounts for the ozone transport of the observed atmosphere; "Prescribed O3 + Adiab. Cooling") shows very little sensitivity to upwelling velocity at this range. In the fast upwelling regime, inclusion of adiabatic cooling shows cold point tropopause upwelling sensitivity, which is similar to that when both effects are accounted for, while inclusion of only ozone transport does not show much cold point tropopause sensitivity. Therefore, ozone transport is the dominant effect in the slow upwelling regime, and adiabatic cooling is the dominant effect in the fast upwelling regime. On the observed order of magnitude of upwelling (which resides between the two regimes) exclusion of either adiabatic cooling or ozone transport reduces the cold point sensitivity to upwelling substantially, compared to when both effects are included.

The cold point temperature results of Figure 2 are mirrored in the cold point altitudes; colder cold points are higher while warmer cold points are lower. One clear pattern from the equilibrium altitude results is that each model version has some upwelling velocity at which the cold point altitude is highly sensitive to upwelling changes. These upwelling values all occur when a near-isothermal region (similar to the "Ozone Trans." case of Figure 1) exists in the temperature profile. Under these conditions, the cold point tropopause is only weakly defined, and therefore, rapid cold point lofting (descent) occurs with only a small increase (decrease) in upwelling.

4. TTL Sensitivity to Transient Upwelling

BDC upwelling in the real atmosphere is variable, due to a seasonal cycle as well as shorter-term variability in wave forcing. To examine the impact of changes in upwelling, we applied perturbations to upwelling, which were vertically uniform and sinusoidally varying in time with periods ranging from 10 to 1/100 years. In all cases the model cold point temperature had a sinusoidal response, which we filtered to select only the mode of the upwelling forcing period (i.e., to remove naturally arising oscillations). Perturbation of only the upwelling for adiabatic cooling produced a larger response amplitude than perturbation of only ozone transport upwelling, while the sum of these amplitudes approximately equaled that when both upwelling velocities were perturbed. We therefore quantified the contribution of the ozone transport-radiation feedback as the ratio of response amplitudes when only upwelling for ozone transport was perturbed compared to when both upwelling velocities were perturbed. This contribution is shown in Figure 3.

The strongest sensitivity of the contribution to forcing period is found around 0.2 years (73 days), similar to the estimated TTL radiative relaxation time scale (Hitchcock et al., 2010). As the ozone transport-radiation feedback operates through the radiative response of the TTL, this similarity is no surprise, and in fact, this represents a method for estimating the TTL radiative relaxation time scale. Below this time scale, the ozone transport-radiation feedback contribution approaches zero as radiation does not have enough time to respond to changes in ozone. Meanwhile, above this time scale there is ample time for a radiative response to ozone changes, and the ozone transport-radiation feedback effect saturates around time scales of half a year, reaching a maximum of about 20%. The contribution slightly decreases at periods longer than 2 years, which could be related to the measure of cold point sensitivity, smaller magnitudes of naturally occurring cold point variation at longer periods, or unidentified mechanisms in the model. Because the decrease is slight, we have not investigated it further.

Furthermore, we found a similar contribution by a second estimate, where the cold point response to a constant forcing of upwelling was examined. In this case the response was clearly linear after about half a year (not shown), and the sum of sensitivities from forcing of only one upwelling velocity nearly equaled that of forcing both upwelling velocities. These sensitivities were constant with upwelling forcing rate, from miniscule to massive forcing rates (0.0025% to 20% per decade, including the expected 2–3% per decade from climate model results). Precisely, the sensitivities were 38.1, 31.34, and 6.83 K/(mm/s), under forcing of both upwelling velocities, only the adiabatic cooling upwelling, and only the ozone transport upwelling, respectively, suggesting an 18% contribution of the ozone transport-radiation feedback.

5. Discussion

Previous RCE studies have shown that ozone has an important influence on the temperature structure of the TTL (BC17; Dacie et al., 2019; McElroy et al., 1992; Thuburn & Craig, 2002). However, these results all used prescribed ozone profiles and explored the TTL temperature sensitivity to ozone under arbitrary ozone changes and therefore could only speculate on the effect of ozone-controlling processes to the TTL temperature structure. This work investigated this subject in greater detail by the inclusion of interactive ozone accounting for simplified stratospheric chemistry and vertical transport into an existing RCE model (BC17). Because of the high level of idealization, it is possible to apply different upwelling velocities to ozone transport and adiabatic cooling within this model (see section 2 for details) and perform a large number of sensitivity experiments. Our results showed that the model is capable of producing the observed structure of the tropical temperature and ozone profiles, despite accounting for a relatively small number of processes (see Figure 1). Of particular interest to this study is the indirect effect of BDC upwelling on temperatures through its control on ozone, referred to in this paper as the *ozone transport-radiation feedback*. This feedback reinforces the impacts of changes in upwelling—an increase in upwelling cools temperatures directly through adiabatic cooling and reduces local ozone concentrations, which also reduces shortwave heating rates and thereby cools the profile further, and vice versa for decreases in upwelling.

In an equilibrium state (Figure 1) our results support those of earlier single-column RCE studies showing that upwelling is of first-order importance for tropical temperature structure via its direct effects through adiabatic cooling (BC17; Dacie et al., 2019). These results also show the well-known first-order influence of upwelling on ozone (Abalos et al., 2013; Avallone & Prather, 1996; Solmon et al., 1985). Furthermore, our results clearly demonstrate that BDC ozone transport has a first-order influence on TTL temperatures through the ozone transport-radiation feedback. In particular, exclusion of BDC ozone transport produced equilibrium lower-stratospheric ozone that was an order of magnitude more dense than observations and a temperature profile that was tens of Kelvin warmer than observed and structurally unlike observations.

Our results also show significant contribution of the ozone transport-radiation feedback to cold point tropopause temperature sensitivity in response to upwelling perturbations. This contribution ranges in strength from about 20% on long BDC forcing periods and decreases toward 0% on shorter time scales. The long-time scale feedback contribution was also estimated with a separate experiment by applying constant upwelling accelerations and the contribution in this case was also about 20%, which was constant with respect to the upwelling trend. The feedback contribution showed little sensitivity to an increase in the BDC forcing period above time scales of about half a year, but large sensitivity on shorter time scales. This is due to the radiative mechanism of the feedback, as upwelling perturbations on time scales shorter than the TTL radiative relaxation time scale do not provide sufficient time for TTL temperatures to respond. However, on time scales longer than the TTL radiative relaxation time scale, the TTL has ample time to adjust to changed ozone concentrations and the feedback effect is saturated.

In general, these results show that perturbations of BDC upwelling primarily affect TTL temperatures through adiabatic cooling but that the ozone transport-radiation feedback can play a significant role as well. This is broadly consistent with prior investigations of the influence of ozone in the TTL temperature cycle (Fueglistaler et al., 2011; Ming et al., 2017). These results also suggest that the ozone transport-radiation feedback can play a role not only for the impacts of climate change, the El Niño southern oscillation, and the quasi-biennial oscillation but also for shorter-time scale phenomena such as the Madden-Julian oscillation. Furthermore, although the ozone transport-radiation feedback provides a minority of the total impact of upwelling, this contribution could have outsized impacts via the effect of tropical tropopause freeze drying in affecting stratospheric water vapor content. This work also presents a possible method for calculation of the TTL radiative relaxation time scale by identification of the BDC forcing period showing the greatest ozone transport-radiation feedback contribution sensitivity. In this case, we find a radiative relaxation time scale of about 70 days, which is in agreement with previous estimates (Hitchcock et al., 2010).

In closing, we reemphasize that the model in this work is idealized and that we have isolated particular processes, while neglecting others. For example, we have not considered variations in surface temperature, which are known to have a strong impact on tropopause temperatures (e.g., Austin & Reichler, 2008; Dacie et al., 2019), or cloud radiative effects, which may have significant effects within the TTL (Fu et al., 2018). Nevertheless, for the purpose of improved understanding, the isolation of particular processes is crucial (the ozone transport-radiation feedback in this work), underlining the importance of employing idealized models in addition to comprehensive models.

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