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Key Points:
- Future lower stratospheric temperature trend is mainly due to GHG increases
- Radiative effects of ozone changes by vertical upwelling are the main cause
- The dynamical cooling by vertical upwelling is a minor contributor

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Seasonality in future tropical lower stratospheric temperature trends

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Abstract The seasonality of the 21st century trends in tropical lower stratospheric temperature (LST) is examined in simulations by a group of comprehensive chemistry-climate models. In contrast to the past LST trends, there is robust seasonal dependence among ensembles of the same model. Furthermore, most models show strongest cooling around July–September and minimal cooling in February–March, which results in a weakening of the seasonality in tropical LST. Sensitivity simulations with isolated forcing reveal that greenhouse gas increases dominate the future tropical LST trend. This seasonally varying LST trend is linked to changes in the Brewer-Dobson circulation (BDC). The BDC can influence the LST through direct dynamical heating/cooling and indirect radiative effects primarily from ozone changes due to vertical transport. The latter is found to be the main cause for the seasonality of the 21st century LST trend, while it is difficult to separate them in the past.

1. Introduction

The lower stratospheric temperature (LST) plays an essential role in troposphere-stratosphere exchange and thus is key to stratospheric chemistry [e.g., *Plumb*, 2002; *Randel et al.*, 2009; *Seidel et al.*, 2011]. The LST is controlled by the coupled effects of dynamical, radiative, and chemical processes, and the LST trend can help understand these dynamical and radiative processes and how they respond to anthropogenic forcing. The observed and simulated LST trends in the past few decades have shown significant seasonal variations in all latitudes [e.g., *Thompson and Solomon*, 2005; *Fu et al.*, 2010; *Free*, 2011]. On the other hand, the seasonality of the simulated LST trends varies largely among models and even among ensemble members that only differ in their initial conditions and in general does not match the observed seasonality [*Wang and Waugh*, 2012].

Despite the nonrobustness in the seasonality, a seasonal anticorrelation between the tropical and extratropical LST trends can be found in most simulations, in agreement with the observations [*Wang and Waugh*, 2012]. This anticorrelation has been linked to the changes in the Brewer-Dobson circulation (BDC) [e.g., *Fu et al.*, 2010], similar to the mechanism for the seasonality in the climatological LST proposed by *Yulaeva et al.* [1994] and supplemented by *Randel et al.* [2002, 2007] and *Fueglistaler et al.* [2011]. In addition to causing direct adiabatic cooling, a stronger BDC (tropical upwelling) brings more ozone-poor air from below and thus induces a negative trend in the lower stratospheric ozone (LSO_3) that leads to indirect diabatic cooling in the tropical lower stratosphere and vice versa for a weaker BDC. Similarly, the sinking branch of the BDC in the extratropics causes an opposite effect on the LST trends, which can explain the anticorrelation between the two regions.

There is still ongoing debate on how the BDC drives the seasonality of the LST trends. In particular, it remains unclear whether the link between the seasonality of trends in tropical LST and upwelling is due to direct changes in adiabatic cooling or due to in situ radiative effects by changes in ozone resulting from the seasonality in tropical upwelling trends. Several studies indicate that the radiative effects due to decreases in ozone are the major factor in temperature trends in the tropical lower stratosphere over the last half century [e.g., *Cordero and Forster*, 2006; *Forster et al.*, 2007; *Polvani and Solomon*, 2012]. The seasonality of LST trends matches that observed in simulations where LSO_3 is prescribed from observations [*Polvani and Solomon*, 2012], but in most models with interactive ozone, the seasonality of LST and LSO_3 does not match that observed [e.g., *Wang and Waugh*, 2012].

Although several studies have investigated the seasonality of LST trends over the past few decades, the seasonality of future LST trends has not been examined. Here we study this seasonality in simulations from a

Table 1. List of CCMVal-2 Simulations[a]

Models	Number of Ensembles		
	GHG + ODS	GHG only	ODS only
CAM3.5 (Community Atmosphere Model 3.5)	1	0	0
CCSR-NIES (Center for Climate System Research–National Institute of Environmental Studies)	1	1	1
CMAM (Canadian Middle Atmosphere Model)	3	3	3
CNRM-ACM (Centre National de Recherches Météorologiques–ARPEGE Climate coupled MOCAGE)	1	0	0
GEOSCCM (Goddard Earth Observing System chemistry-climate model)	1	1	0
LMDZrepro (Laboratoire de Météorologie Dynamique Zoom–REPROBUS)	1	1	1
MRI (Meteorological Research Institute)	2	1	1
NIWA-SOCOL (National Institute of Water and Atmospheric Research–Solar Climate Ozone Links)	1	0	0
SOCOL (Solar Climate Ozone Links)	3	1	1
UMSLIMCAT (Unified Model–SLIMCAT)	1	1	1
UMUKCA-METO (Unified Model/UK Chemistry and Aerosols Module–Met Office)	1	0	0
UMUKCA-UCAM (Unified Model/UK Chemistry and Aerosols Module–University of Cambridge)	1	1	0
WACCM (Whole-Atmosphere Chemistry-Climate Model)	3	1	1

[a]See *Morgenstern et al.* [2010] for details.

suite of chemistry-climate models (CCMs). We analyze the seasonality in the 21st century LST trends and compare with the past trends in the same models. We examine the magnitude and phasing of the seasonality, the robustness of the seasonality among simulations, and the cause of the seasonality. It is shown that there is large seasonality in the future LST trends that is linked to increases in greenhouse gases (GHGs) and the seasonality in upwelling. Furthermore, it is shown that in situ radiative effects due to changes in ozone, and not the direct adiabatic cooling, are the primary mechanism causing the seasonality of LST trends.

The manuscript is organized as follows. Section 2 describes the data and models used and methods of our analyses. In section 3, we focus on the Canadian Middle Atmosphere Model (CMAM) [*de Grandpré et al.*, 2000; *Scinocca et al.*, 2008, 2009] to analyze the tropical LST trend and the dynamics controlling its seasonality. Then we examine the findings from CMAM in other model simulations in section 4, followed by the conclusion and discussion.

2. Data and Methods

We examine here the output from chemistry-climate model simulations done as part of the Stratospheric Processes and their Role in Climate (SPARC) Chemistry-Climate Model Validation (CCMVal) Activity phase 2 (referred to as CCMVal-2) [*Stratospheric Processes and their Role in Climate Chemistry-Climate Model Validation*, 2010]. As part of CCMVal-2 simulations, three different scenarios were performed: a REF-B2 simulation in which both ozone depleting substances (ODSs) and GHGs are varied with time, a SCN-B2b simulation where only GHGs varied, and a SCN-B2c simulation where only ODSs varied. These three scenarios are referred to below as "GHG + ODS," "GHG only," and "ODS only," respectively. The GHG boundary conditions used follow the *Intergovernmental Panel on Climate Change* [2001] Special Report on Emissions Scenarios A1b scenario, while the ODSs follow the adjusted *World Meteorological Organization* [2007] A1 scenario (see *Eyring et al.* [2008] and *Morgenstern et al.* [2010] for more details). A full list of the CCMVal-2 model simulations can be found in Table 1.

We focus here mostly on simulations by CMAM. This is the only CCM for which there are multiple ensemble members of each of the three scenarios (GHG + ODS, GHG only, and ODS only), and all simulations run from 1960 to 2099. Furthermore, CMAM includes an interactive dynamic ocean [e.g., *Scinocca et al.*, 2009; *McLandress et al.*, 2010, 2011]. The remaining CCMVal-2 models use prescribed sea surface temperatures (SSTs) from coupled atmosphere-ocean models. CMAM also performs well in the comparison with observations of stratospheric dynamics and transport [*Butchart et al.*, 2011; *Strahan et al.*, 2011].

The model fields examined include the zonal mean temperature, ozone, and transformed Eulerian mean residual vertical velocity (w^*). Monthly mean values at 70 hPa averaged over 20°S–20°N are used to calculate the LST/O_3/w^* trends. In our analysis, we divide the data into a past or ozone depletion period (1960–1999) and a future or ozone recovery period (2000–2099). Note that the magnitude of the decrease from 2000 to 2099 is similar to that of the 1960 to 2000 increase in ODSs. Linear trends over these periods are calculated

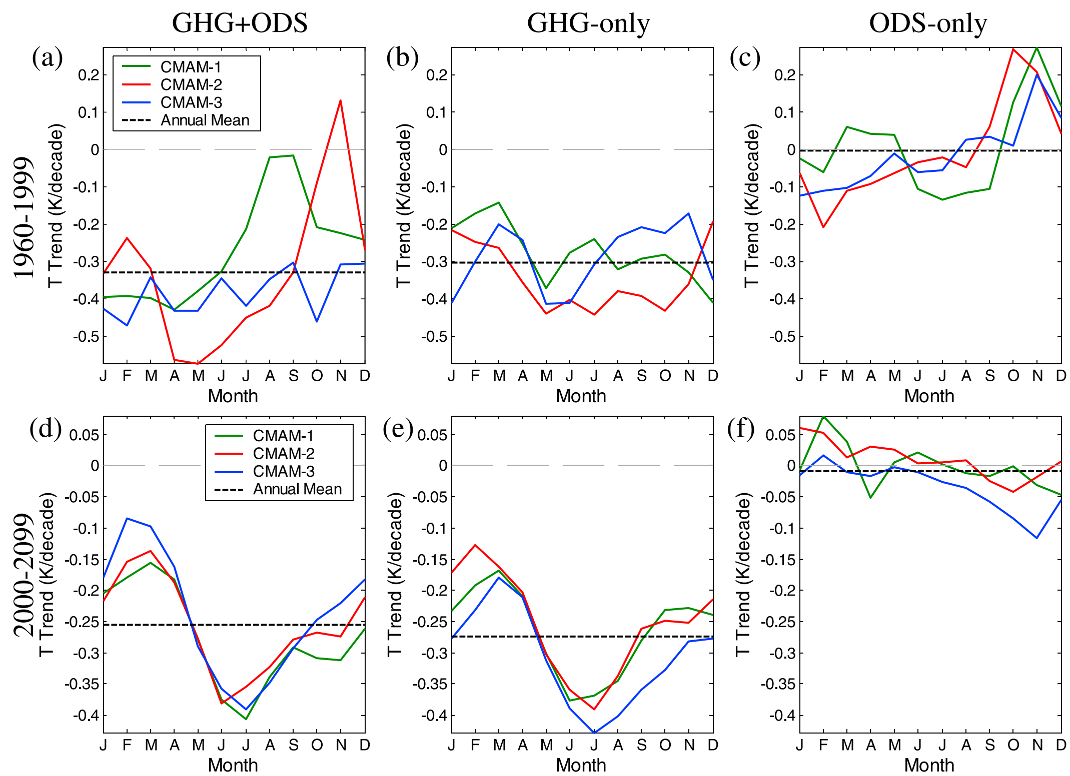

Figure 1. The 1960–1999 monthly trends in CMAM tropical (20°S–20°N) lower stratospheric (70 hPa) temperature in response to changes of (a) both GHGs and ODSs, (b) GHGs only, and (c) ODSs only. (d–f) Similar to Figures 1a–1c but for 2000–2099 trends. The three colors represent three ensemble members that only differ in their initial conditions. The black dashed line illustrates the ensemble mean annual mean trend in each panel.

using the standard least squares estimation. The tropical lower stratospheric temperature trends greater than 0.05 K/decade are generally statistically significant, with consideration of autocorrelation of the residuals [*Weatherhead et al.*, 1998]. Since the temperature trends are significant for most models in most months, error bars are not shown on the figures for clarity.

The connection between the seasonally varying trends in different fields and the robustness of seasonality among simulations is determined by calculating linear correlations. The statistical significance of the linear correlation can be quantified by the Student's *t* test. The number of degree of freedom is the number of data points minus 2 (because the correlation involves two members; 12–2 = 10 for monthly trends), and the significance thresholds for the correlation coefficients are 0.497 at the 10% level, 0.576 at the 5% level, and 0.708 at the 1% level. The robustness of the seasonality in the LST trends is evaluated by averaging seasonal correlation of all pairs of ensemble members (for example, three pairs for three ensembles).

3. CMAM Simulations

3.1. Tropical LST Trends

We first consider the LST trends in the CMAM. Three ensemble members of each of the scenarios (GHG + ODS, GHG only, and ODS only) are available from CMAM, and we can examine the robustness in LST trends among ensemble members as well as the relative role of changes in ODSs or GHGs in driving the trends.

Wang and Waugh [2012] showed that the seasonality of the 1960–1999 LST trends is not robust across ensemble members of the CMAM REF-B1 simulations, which use observed, instead of projected, forcing. This nonrobustness in the seasonality of the past (1960–1999) trends among ensemble members is also found for the CMAM GHG + ODS simulations during the same period, see Figure 1a. However, as shown in Figure 1d, there is a robust seasonality in the future (2000–2099) LST trends in the GHG + ODS simulations. All three ensemble members show large seasonality in the 2000–2099 trends, with strongest cooling around July

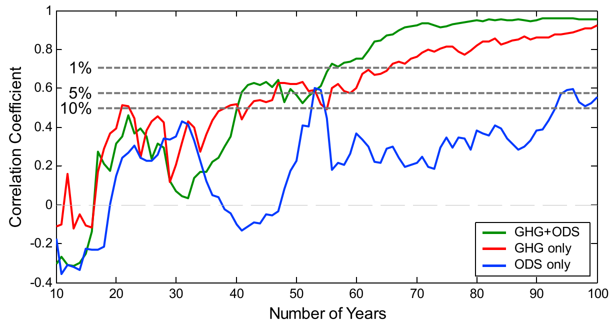

Figure 2. Averaged correlation between ensemble members of CMAM tropical LST trends, in response to changes of both GHGs and ODSs (green), GHGs only (red), and ODSs only (blue), over different number of years in 2000–2099. The statistical significance is labeled at 1%, 5%, and 10% by *t* test.

(−0.4 K/decade) and weakest cooling around February (−0.1 K/decade). This contrast in the robustness in seasonality between past and future trends also holds for the sensitivity simulations where only GHGs vary with time, see Figures 1b and 1e. For both GHG + ODS and GHG-only scenarios, there is a large spread among ensemble members for 1960–1999 trends but similar seasonality for the 2000–2099 trends.

The main cause for the difference in robustness among ensemble members between past and future trends is the longer time period for the future trends (100 years compared with 40 years). If only a period of 40 years (e.g., 2000–2040) is used to calculate future trends, then there are also differences among ensemble members. The dependence on length of time used for the trend is quantified in Figure 2, which shows the average correlation between different ensemble members for trends of different lengths. This indicates that only when 70 years or longer is used, there is convergence in the seasonality of future trends for the GHG + ODS and GHG-only simulations. For ODS-only simulations, there is still a spread among the ensemble members for 100 year trends, but as shown in Figure 1f, the seasonality is small for all members. Note that the direct chemical impact of ODSs on O_3 in the tropical lower stratosphere is very small. The LST response seen in Figures 1c and 1f is primarily dynamical through w^* that is indirectly affected by ODS changes. For example, past ODS increases induce severe ozone depletion in the polar stratosphere, which strengthens the polar vortex and then increases extratropical wave activity to enhance tropical upwelling [*Fu et al.*, 2010; *Wang and Waugh*, 2012].

Comparison of the trends among the different scenarios (different columns in Figure 1) shows that the annual mean LST trend (horizontal dashed lines in Figure 1) is primarily explained by increases in GHGs rather than changes in ODSs; i.e., the annual mean trends in ODS simulations are near zero, while the annual mean trends for GHG + ODS and GHG simulations are similar, for both the past and future trends. However, the cause of the seasonality of the trends differs between the past and the future: the seasonality of the 1960–1999 LST trend depends on both the GHG and ODS changes (Figures 1a and 1e), whereas the seasonality in the 2000–2099 LST trend primarily comes from the GHG increases (Figures 1b and 1d). This difference occurs because the amplitude of the seasonality in trends due to ODSs is much smaller in the future than in the past (which is related to slower rate of ozone recovery compared to ozone depletion). The magnitude of the seasonality of the trends due to GHG is similar over the two periods and is comparable to that due to ODSs in the past but much larger than that due to ODSs in the future.

In sum, all three ensemble members of CMAM show large seasonality in the 2000–2099 LST trends, with strongest cooling around July and weakest cooling around February (which results in a weakening of the seasonality in LST). This agreement among ensemble members only occurs for trends over 70 or more years, and there is not a robust seasonality of LST trends for the past (1960–1999) or the first half of the 21st century (2000–2050). The dominant forcing of the seasonality of LST trends varies between the past and the future. The seasonality of the past LST trends is driven by both increases in ODSs and GHGs, whereas increases in GHGs are the main driver of the seasonality of future LST trends.

3.2. Mechanisms

We now examine the mechanisms causing the seasonality in future LST trends. Temperature and ozone trends in the tropical lower stratosphere are closely related together as ozone is the major absorber of incoming short-wave solar radiation and an effective absorber of outgoing long-wave radiation. This close connection occurs in all three scenarios, where the LSO_3 and LST trends have very similar seasonality, see Figure 3. The ensemble mean trends are shown in Figure 3, but the seasonality of trends in LSO_3 for individual ensemble members is also similar to that for LST; e.g., there is a robust seasonality for future GHG + ODS trends but not for past GHG + ODS trends.

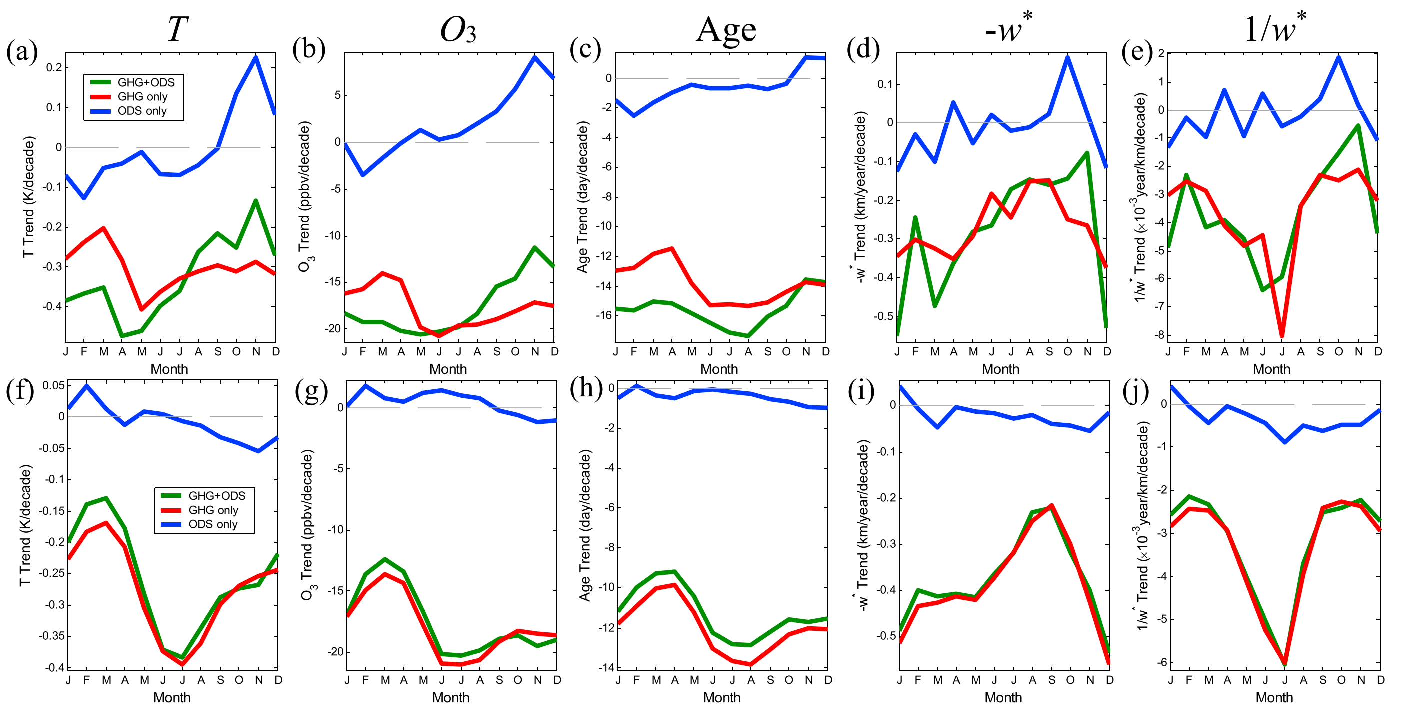

Figure 3. The 1960–1999 monthly trends in CMAM ensemble mean tropical (20°S–20°N) lower stratospheric (70 hPa) (a) temperature, (b) ozone, (c) mean age of air, (d) $-w^*$, and (e) $1/w^*$, in response to changes of both GHGs and ODSs (green), GHGs only (red), and ODSs only (blue). (f–j) Similar to Figures 3a–3e but for 2000–2099 trends.

As discussed in the Introduction, the seasonality of the observed (and simulated) LST trends has been related to the seasonality of the trends in tropical upwelling. Changes in upwelling can impact LST both by direct adiabatic cooling (i.e., increased upwelling results in adiabatic cooling and a decrease in LST) and indirectly through the radiative effects due to vertical ozone transport. An increase in upwelling brings more ozone-poor air from below and induces a negative trend in the lower stratospheric ozone, and this decrease in ozone then results in less diabatic heating and a negative trend in LST. It is generally difficult to separate the direct role of adiabatic cooling and the indirect role ozone transport as there is a high seasonal correlation between LST and LSO_3 trends (e.g., Figure 3). However, consideration of the first-order balances for T and O_3 in the tropical lower stratosphere suggests that LST and LSO_3 may have different relationships with w^*.

The approximate thermodynamic balance in the tropical lower stratosphere can be written as [e.g., *Randel et al.*, 2002]

$$\frac{\partial T}{\partial t} + w^* S = -\alpha (T - T_{EQ}) \tag{1}$$

where T is the zonal mean temperature, w^* is the residual vertical velocity, S is the static stability, α is an inverse radiative damping time scale, and T_{EQ} is the radiative equilibrium temperature. (We have dropped the usual convention of an overbar for zonal mean quantities as all quantities are zonal mean values.) If we assume that T is approximately steady state and changes in S, α, and T_{EQ} are small, we have $T = -w^* S/\alpha$, and we would expect seasonal variations of trends in T to be correlated with trends in $-w^*$. It is reasonable to assume that small changes in the static stability as the stratospheric static stability is primarily maintained by the radiative heating from the entire column of the atmosphere [e.g., *Young*, 2003]. Furthermore, $S = \partial T/\partial z + g/c_p$ (g is the acceleration of gravity and c_p is the specific heat of air), and $\partial T/\partial z$ is usually an order of magnitude less than g/c_p in the stratosphere, so S is not sensitive to changes in $\partial T/\partial z$. For example, the annual mean climatology of $\partial T/\partial z$ is reduced by over 30% from the first decade to the last decade of 21st century in the CMAM GHG + ODS simulation, but S decreases by less than 10%. In contrast to the large annual cycle in T and w^*, S has a weak semiannual cycle in the current climatology and varies within 3.5% of its annual mean.

For O_3, the balance in the tropical lower stratosphere is somewhat different. To leading order, the zonal average continuity equation for O_3 mixing ratio is

$$\frac{\partial O_3}{\partial t} + w^* \frac{\partial O_3}{\partial z} = P \quad (2)$$

where P is the net ozone production, and we have ignored horizontal advection and eddy transport (dominated by meridional mixing, which is approximately proportional to the ozone mixing ratio difference between the tropics and extratropics [e.g., *Monier and Weare*, 2011]). It is not possible to assess the impact of the neglect of the eddy transport, as the required model output has not been archived. The possible implications of neglecting horizontal mixing are discussed in section 4. Further, assuming steady state and small changes in P, we have $\partial O_3/\partial z = P/w^*$. P in the tropical lower stratosphere is primarily affected by the short-wave solar radiation and thus only has a weak semiannual cycle due to variations in solar zenith angle (not shown). Since the tropospheric ozone mixing ratio is much smaller than its stratospheric counterpart, $\partial O_3/\partial z \approx O_3/\delta z$ (δz is the altitude relative to the tropopause) is a good approximation in the tropics for a small distance above the tropopause. The $\partial O_3/\partial z$ thus cannot be assumed constant because of this tight connection to local ozone changes and its distinct annual cycle similar to O_3 (not shown). Therefore, $O_3 \approx P\delta z/w^*$ and a positive correlation would be expected between trends in O_3 and $1/w^*$.

The O_3 balance in the tropical lower stratosphere is very similar to that for the mean age of air (referred to as "age" below for simplicity). In the nondiffusive limit, which is a good approximation in the tropics [e.g., *Neu and Plumb*, 1999; *Hall and Waugh*, 2000], the age satisfies

$$\frac{\partial \Gamma_T}{\partial t} + w^* \frac{\partial \Gamma_T}{\partial z} = 1 - \mu(\Gamma_T - \Gamma_{ET}), \quad (3)$$

where Γ_T and Γ_{ET} are the tropical and extratropical age and μ is a parameter that depends on the ratio of mass between the tropics and the extratropics, the effective rate of net entrainment constrained by mass flux and horizontal mixing. There exists an analytical solution for vertically uniform w^*, where $\Gamma_T \propto \delta z/w^*$ regardless of the presence of horizontal mixing [*Neu and Plumb*, 1999; *Hall and Waugh*, 2000; *Garny et al.*, 2014]. So given the similarity between equations (2) and (3), a positive correlation would be expected between trends in O_3 and $1/w^*$. The seasonality in LSO_3 and age trends are almost identical (Figure 3), indicating the close similarity in their response to changes in w^*.

A time lag of 1 month or so is expected in the response of O_3 and age to w^* due to the vertical transport [e.g., *Chae and Sherwood*, 2007]. For example, w^* is around 20 km/yr in the tropical lowermost stratosphere, corresponding to a 1.5 months advection time (or age at the no horizontal mixing limit) from 100 hPa to 70 hPa ($\delta z \approx 2.5$ km).

The above reasoning shows fundamental differences in the relationship of T and O_3 with w^* in the lower stratosphere. It suggests that if trends in w^* are directly driving trends in T, we would expect the seasonality of T trends to be correlated with seasonality of trends in $-w^*$, whereas if w^* drives trends in O_3, which then causes changes in T_{EQ} and hence changes in T (because $T = T_{EQ} - w^*S/\alpha$ when changes in T_{EQ} are no longer small), trends in T would be correlated with $1/w^*$ because of the positive dependence of T_{EQ} on O_3. So looking at the correlations between the seasonality of trends in T and $-w^*$ or $1/w^*$ can potentially isolate the mechanism driving the seasonality in LST trends.

Figure 3 shows the seasonality of $-w^*$ and $1/w^*$ for the past and future trends ($-w^*$ is plotted rather than w^* because of the above expectation of a positive correlation between T and $-w^*$). For the past GHG + ODS simulations, the seasonality of $-w^*$ and $1/w^*$ is somewhat similar, and the seasonal correlations $r(T, -w^*)$ and $r(T, 1/w^*)$ are of similar magnitude (see Figure 4a). Thus, for this simulation, it is not possible to use this metric to separate the mechanisms involved. This also holds for both the past and future ODS-only simulations, where the seasonality of $-w^*$ and $1/w^*$ as well as correlations with T are similar.

The situation is different for the future GHG + ODS simulation and both the past and future GHG-only simulations. For these simulations, the seasonality of $-w^*$ and $1/w^*$ trends and their correlations with T are very different, see Figures 3 and 4. In particular, the correlation between T and $-w^*$ is less than zero, which is the opposite sign to that expected if the seasonality in LST trends were due to variations in adiabatic heating. In contrast, there are still positive correlations between T and $1/w^*$ (with or without the 1 month lag by vertical transport). This

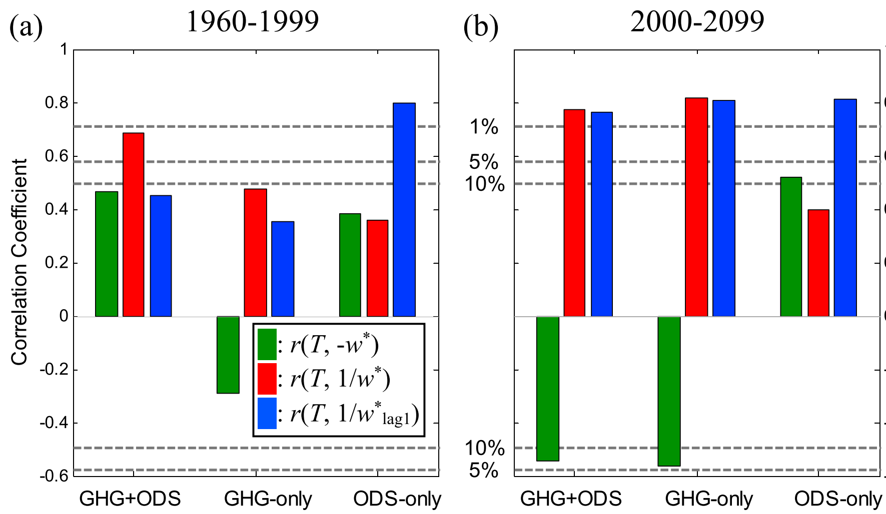

Figure 4. Seasonal correlation (r) of CMAM ensemble mean tropical (20°S–20°N) lower stratospheric (70 hPa) trends between temperature and $-w^*$ (green), $1/w^*$ (red), and $1/w^*$ with 1 month lag (blue), for the period of (a) 1960–1999 and (b) 2000–2099, in response to changes of both GHGs and ODSs, GHGs only, and ODSs only. The statistical significance is labeled at 1%, 5%, and 10% by t test.

suggests that the changes in adiabatic cooling are not the primary cause of the seasonality of the future T trends, and that changes in the vertical advection of O_3, and resulting changes in diabatic heating, are the dominant mechanism. The close similarity in LST, LSO_3, and age trends also supports this inference.

Note that the correlations between O_3 trends and $-w^*$ or $1/w^*$ are generally similar to those of T, i.e., negative correlations between O_3 and $-w^*$ and positive correlations between O_3 and $1/w^*$ for the future GHG + ODS simulation (not shown).

It is of interest to consider the impact of the above seasonally varying LST trends on the seasonality of LST. There is an increase in the annual cycle in the tropical upwelling, while the annual cycle amplitudes of lower stratospheric temperature and ozone are both reduced in the future (Figure 5). The annual cycle amplitude decrease in LSO_3 and increase in w^* are consistent with $O_3 \propto 1/w^*$. The decrease in LSO_3 annual cycle weakens the LST annual cycle amplitude, consistent with the findings that a significant portion of the observed LST annual cycle is contributed by LSO_3 [e.g., *Chae and Sherwood*, 2007; *Fueglistaler et al.*, 2011]. However, this contradicts to the classic view of linear dependence of LST annual cycle amplitude on that of w^* [e.g., *Randel et al.*, 2002], most likely because the background equilibrium temperature T_{EQ} is altered by the LSO_3 long-term

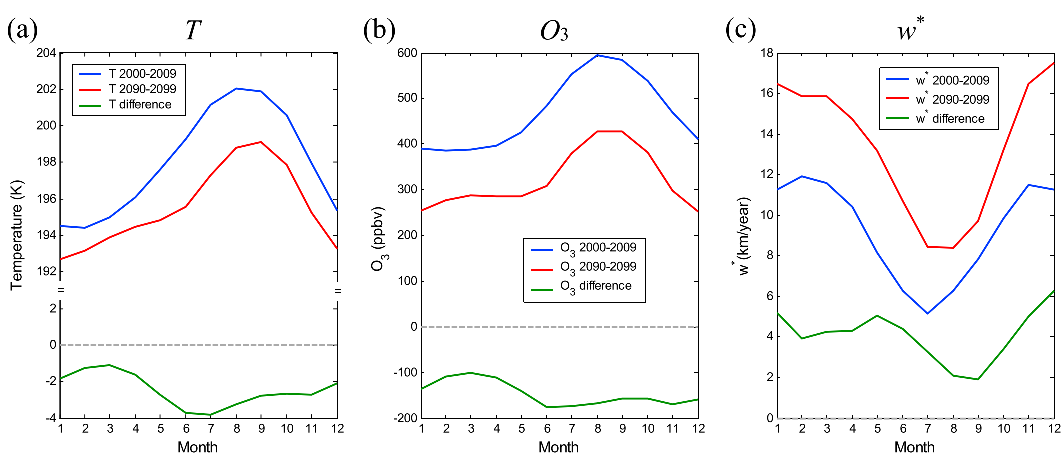

Figure 5. CMAM ensemble mean monthly climatology in tropical (20°S–20°N) lower stratospheric (70 hPa) (a) temperature, (b) ozone, and (c) w^* for the periods of 2000–2019 (blue) and 2080–2099 (red), along with their difference (green) as response to changes of both GHGs and ODSs.

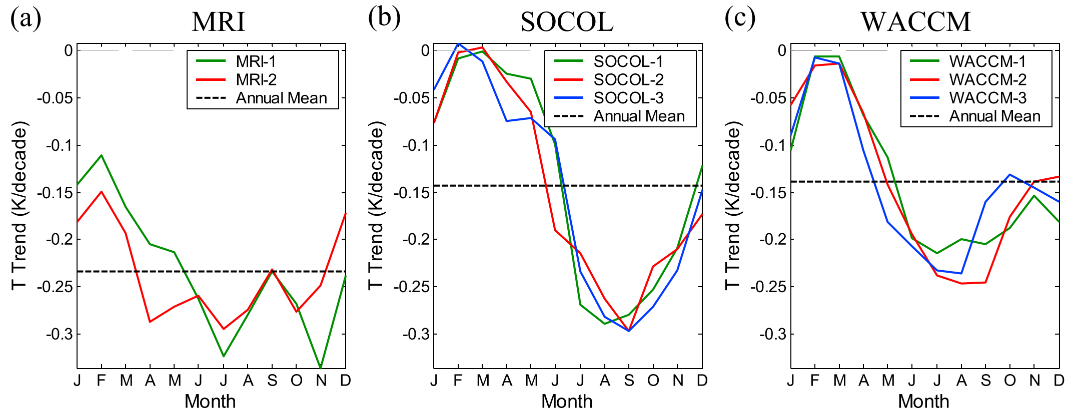

Figure 6. The 2000–2099 monthly tropical (20°S–20°N) lower stratospheric (70 hPa) temperature trends in (a) MRI, (b) SOCOL, and (c) WACCM, in response to changes of both GHGs and ODSs.

changes. Therefore, the w^* seasonality is still the main driver for the annual cycle in LST and LSO_3 for each climatological period, but differences should be noted when comparing two well-separated periods under GHG forcing.

4. CCMVal-2 Models

The above analysis has focused on a single CCM, and hence, the results may be model dependent. To evaluate this model dependence, we examine simulations from other CCMs that participated in the CCMVal-2 activity. Although CMAM is the only model for which multiple ensemble members of each of the three scenarios are available in the CCMVal-2 archive, a few other models performed multiple ensembles of the GHG + ODS simulation or performed single GHG-only and ODS-only simulations. Simulations from these models can be used to examine whether (i) the robustness among ensemble members and (ii) the relative role of GHGs and ODSs in forcing changes in LST found in CMAM hold in other models.

Multiple GHG + ODS simulations from the Whole-Atmosphere Chemistry-Climate Model (WACCM), Solar Climate Ozone Links (SOCOL), and Meteorological Research Institute (MRI) models are available in the CCMVal-2 archive and show the same robustness among ensemble members as CMAM. For each model, the ensemble members show the same seasonality of future (2000–2099) LST trends (Figure 6), but there are differences among the ensembles for the past (1960–1999) trends or future trends less than around 70 years (not shown). The magnitude of the annual mean trends and the annual amplitude of the future trends vary among these models, but they all show a similar seasonal phase with minimum cooling in northern hemisphere (NH) spring and maximum cooling in NH fall (Figure 6).

Simulations for each of the sensitivity scenarios (ODS only and GHG only) were also performed using the Center for Climate System Research–National Institute of Environmental Studies (CCSR-NIES) and WACCM models. Consistent with the CMAM simulations, the seasonality of the past LST trends in these models is driven by both increases in ODSs and GHGs, whereas increases in GHGs are the main driver of future LST trends, see Figure 7. Also, the seasonality of future LST trends in CCSR-NIES and WACCM is similar to those in CMAM, but there is no agreement among the models in past trends. This lack of agreement among models for past trends is not surprising, given the differences among ensemble members for any individual model.

The SOCOL model also performed simulations for all three scenarios; however, the same changing sea surface temperatures (SSTs) were used for all simulations. These SSTs are from a coupled atmosphere-ocean simulation in which GHGs increase over time, so this indirect impact of increasing GHGs is included in all the SOCOL simulations. The LST trends are similar for all three scenarios, indicating that changes in SSTs driven by increasing GHGs are the main driver of the (seasonality in) future LST trends.

Finally, we consider the LST trends in the GHG + ODS simulations. Figure 8a shows the 2000–2099 LST trends from all CCMs in CCMVal-2. There is a large spread in the annual mean 2000–2099 trends and in the

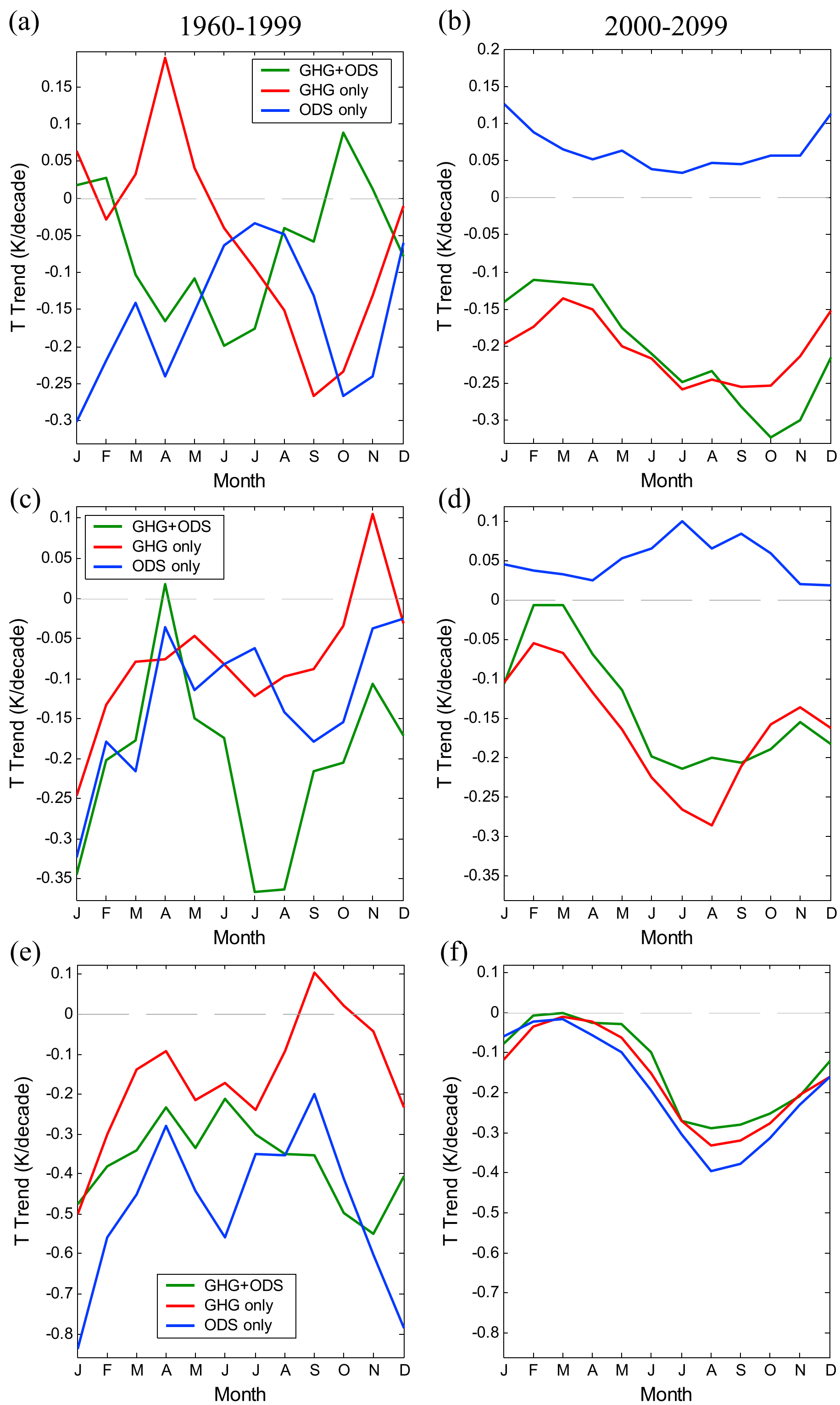

Figure 7. The (a) 1960–1999 and (b) 2000–2099 monthly trends in CCSR-NIES zonal mean tropical (20°S–20°N) lower stratospheric (70 hPa) temperature in response to changes of both GHGs and ODSs (green), GHGs only (red), and ODSs only (blue). (c and d) Similar to Figures 6a and 6b but for WACCM. (e and f) Similar to Figures 6a and 6b but for SOCOL.

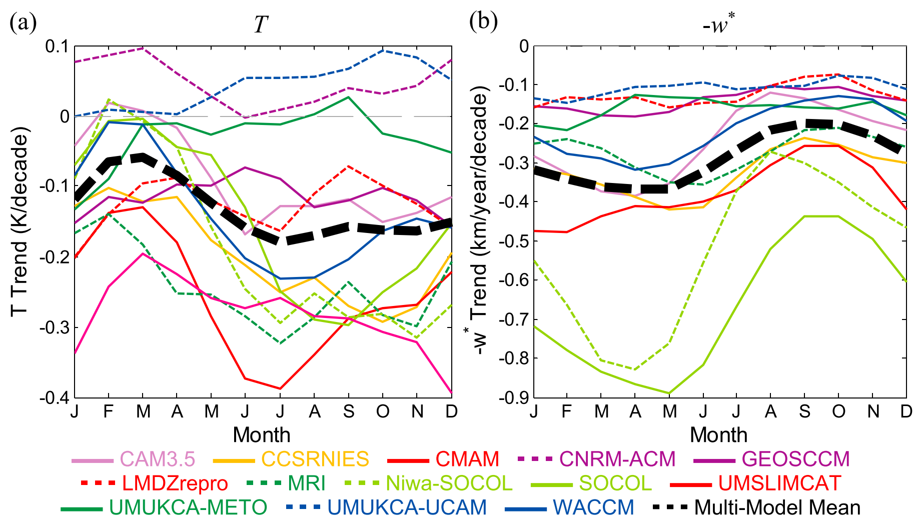

Figure 8. Multimodel 2000–2099 monthly trends in tropical (20°S–20°N) lower stratospheric (70 hPa) (a) temperature and (b) $-w^*$, in response to changes of both GHGs and ODSs. Ensemble mean is shown for models with multiple ensembles. Each model is given the same weight in the multimodel mean regardless of its ensemble size.

seasonality of the trends among the CCMs. Most models show marked seasonal variations in their LST trends, but others show very little seasonality. Figure 8b shows the 2000–2099 w^* trends. The annual cycle amplitude in LST trends is generally consistent with that in their w^* trends.

For simulations showing large seasonality in trends, most models show weakest cooling in February–March and strongest cooling in the late half of the year. This is primarily associated with the seasonality in the trends of vertical ozone advection by w^*. Although the strengthening of tropical w^* is stronger in boreal winter than summer, the changes in $1/w^*$ are the opposite, which lead to the weakest cooling in February–March given the delay by vertical transport. The relative stronger strengthening of w^* has been linked to the hemispheric asymmetry in the subtropical wave breaking in response to GHG increases [e.g., *Shepherd and McLandress*, 2011].

Figure 9 shows the seasonal correlation of future LST trends with $-w^*$ or $1/w^*$ trends (with a 1 month lag due to vertical transport) for the subset of CCMs with a large seasonal variation in w^* trends. There is a large variation in correlation coefficients among the models, but there is a general agreement in the sign. Specifically, $r(T, -w^*)$ is negative or near zero for all models, whereas $r(T, 1/w^*)$ is positive. This supports the above hypothesis that changes in adiabatic cooling are not the primary cause of the seasonality of the future LST trends (as this would result in a positive $r(T, -w^*)$) and that changes in the vertical advection of O_3, and resulting changes in diabatic heating, are the dominant mechanism.

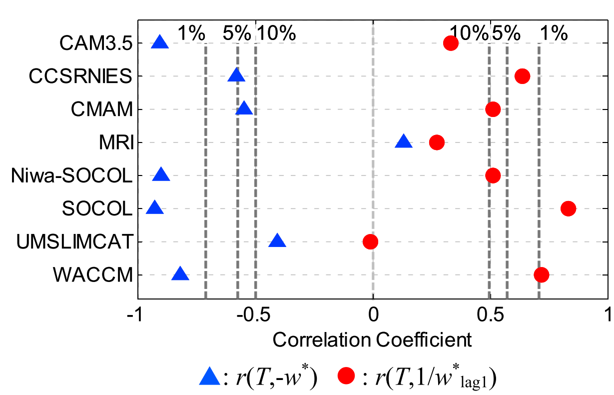

Figure 9. Seasonal correlation of tropical (20°S–20°N) lower stratospheric (70 hPa) temperature trend with $-w^*$ trend (triangles) and $1/w^*$ trend with 1 month lag (dots) in response to changes of both GHGs and ODSs for 2000–2099. The statistical significance is labeled at 1%, 5%, and 10% by *t* test.

Although $r(T, 1/w^*)$ is positive for almost all models, for some, the magnitude is small. This indicates that other processes may also be playing a role, such as changes in horizontal mixing with midlatitudes or ozone production [e.g., *Konopka et al.*, 2009, 2010; *Ploeger et al.*, 2012; *Abalos et al.*, 2012, 2013; *Meul et al.*, 2014]. On one hand, the ozone production rate has weakly temperature dependence and thus may increase in the future because of the cooling in the tropical lower stratosphere. On the other hand, the horizontal mixing might play a role. The subtropical wave breaking not only drives a significant part of the BDC but also contributes to the horizontal mixing. Both can be enhanced by increases in the subtropical wave breaking. An enhanced BDC tends to dilute the LSO_3 and leads to cooling in the tropical lower stratosphere, while stronger horizontal mixing brings more ozone from the extratropics and causes warming in the tropical lower stratosphere. The vertical advection and the horizontal mixing could thus have opposite impacts on the transport. The cooling due to the strengthened vertical ozone transport can be partially cancelled by the enhanced horizontal mixing efficiency in the summer hemisphere, which can vary among models. The climatological horizontal mixing and vertical transport both show large variations among models [e.g., *Strahan et al.*, 2011], which could be a source of uncertainty of the projected response to anthropogenic forcing. Further discussion along this line is beyond the scope of this study.

5. Conclusions

Several key results have been found from our analysis of a suite of CMAM simulations:

1. The seasonality of future LST trends is robust (i.e., there is agreement among ensemble members) for periods of over 70 or more years (e.g., 2000–2070), with strongest cooling around July and weakest cooling around February.
2. Increases in GHGs, and resulting changes in the tropical upwelling, are the dominant cause of the seasonality in future LST trends, with ozone recovery playing only a minor role.
3. The seasonality in tropical upwelling trends causes the seasonality in LST trends primarily by the vertical transport of ozone and resulting changes in radiation absorption by ozone. Direct adiabatic cooling due to the tropical upwelling plays a secondary role.

These results can be contrasted with the seasonality of simulated past LST trends, where there are significant differences between ensemble members, where ozone depletion and GHG increases both play an important role, and where it remains a challenge to quantitatively separate the roles of the direct adiabatic cooling and indirect ozone pathway [e.g., *Fu et al.*, 2010; *Wang and Waugh*, 2012].

The robustness in the seasonality of LST trends over the 21st century among ensemble members is also found for other CCM simulations in the CCMVal-2 archive, as is the dominant role of GHG increases in causing the seasonality of these trends. A majority of these CCMs show large annual cycle amplitudes in their LST trends that have similar seasonality to the CMAM trends. There is also a general agreement in the seasonality of the LST trends forced by GHG increases, which is not found in the ODS-only simulations.

Changes in the BDC are very difficult to observe or measure directly, and as a result, temperature changes have been used to estimate changes in the BDC [e.g., *Young et al.*, 2012]. This approach is based on the anticorrelation between lower stratospheric temperature and the BDC, which is valid in the past but not in the future due to changes in the relative role of direct adiabatic cooling and indirect diabatic cooling by vertical ozone transport. As a result, caution needs to be taken when using the temperature trend to infer the secular seasonal trend in the BDC for the future.

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