Anonymous Referee #1

General Comments

This paper explores the causes behind a recent decadal warming of the tropical tropopause layer (TTL) using a series of well-designed climate model experiments with NCAR's WACCM model. The authors conclude that natural (QBO, SST) variability, rather than anthropogenic factors, are responsible for the recent warming of the TTL. They also illustrate the importance of the model's vertical resolution in simulating temperature trends in the TTL.

The manuscript is generally well written and appropriate for publication in ACP. However, after reading through the manuscript, I have some unanswered questions about the authors' methodology and their interpretation of the results, which I think should be addressed prior to publication.

Specific Comments

1. The warming trend mentioned in this paper from 2001-2011 has not continued to the present (see, e.g., recent article in EOS: DOI: 10.1002/2014EO270001). If you were to redo this analysis with slightly different end years (e.g., 2002 to 2012, 2003 to 2013), you might reach very different conclusions, so I'm not entirely comfortable with the short 11-year trend period used in this paper. The authors need to discuss the sensitivity of their results to the short period chosen for the trend analysis, and make readers aware that the warming trend has not continued (at least monotonically) from 2011 to 2014.

Response:

We thank the reviewer for this comment and confirm that the trends are sensitive to the start and end years in particular for short time series as the one from GPS-RO data investigated here. In order to address this issue, we repeated our analysis with data extending to December 2012, December 2013 and until the latest GPS-RO data available to us, i.e. March 2014 (Fig. S1). In Fig. S2, we also include now the latitude-height structure of the trends for the different time periods. By including more years beyond 2011, the trends are generally weaker (~1.0 K) compared with the trends from 2001-2011 (~1.6 K). The variability of the trend is consistent with our conclusion that the recent decadal TTL variation in temperature is mainly due to natural variability and that this positive trend might not continue so strong in the future.

Changes in manuscript:

We have added two supplementary figures to the paper as well as a paragraph to comment on this in the text.

2. When assessing the statistical significance of trends in time series, autocorrelation can sometimes inflate the significance of a trend. This is easily corrected by changing the n in equation 3 to an effective sample size (see Santer et al. 2000, as well as the Wigley 2006 paper that the authors cite). It's not clear from the text if the authors did this, but if not, they should double check to make sure that their trends are still significant after accounting for autocorrelation.

Thank you very much for this note.

We have reassessed the statistical significances of the trends and now also consider the autocorrelation effects. We added more details of our method in the text and commented the results accordingly.

3. Figures 9b-d and 10c-d are extremely difficult to read. As a reviewer, I cannot properly assess many of the statements of the authors in Sections 4.1 and 4.2 because I cannot see what they are referring to in the figures. Either the signatures the authors are discussing are not robust, or the figures need to be improved. I'm not sure what to suggest here, but I would encourage the authors to closely read their text and make sure their conclusions are readily visible in the figures.

We apologize for that this figure was difficult to read and the conclusions were difficult to follow.

We have combined Figures 9 and 10 and restructured the figure in order to be hopefully better readable now. We've also adapted the corresponding text.

4. Page 22120, Line 25: Reflecting and scattering incident solar radiation back to space does not lead to a warming of the lower stratosphere. Please rephrase.

We have rephrased this sentence to:

"Stratospheric aerosols reflect solar visible radiation, causing a net cooling at the surface, and absorb solar near-infrared and terrestrial infrared radiation, causing a warming of the lower stratosphere, which maximizes at around 20 km."

5. Figure 1c: Please explain how the QBO values are determined in the model for future years.

The QBO forcing time series in CESM-WACCM is determined from a filtered spectral decomposition of the observed climatology from 1953-2004. The resulting set of Fourier coefficients can then be expanded for any day and year into the future. We added an explanation on this to the manuscript (CESM model description).

We have added this explanation in the text.

6. Page 22126, Lines 7-9: Why are the observed SSTs and simulated SSTs (from a fully coupled model) decreasing over exactly the same period? Shouldn't the model's SST variability be internally generated and thus independent from that in the real world?

You are right, the simulated SSTs in the Natural run are internally generated from a fully coupled model. The similar decrease in both model and observations might be entirely due to internal variability of the climate system, or by chance since we have only one ensemble of simulation.

7. Sections 3.2 – Section 3.6: The uncertainty in each figure is listed in each paragraph as 0.2 K/decade. Yet, in some paragraphs, the uncertainty is stated as "small", and in other paragraphs, it's stated as "large." This is confusing. Please clarify and rephrase.

We apologize for not havening explained this clearly. "Small" or "large" means that the uncertainty is relatively small or large compared to the trend itself.

We have changed it accordingly in the text.

Technical Corrections

Page 22119, Line 4: Delete "masses"

Deleted.

Page 22119, Line 19: Delete "for"

Deleted.

Page 22120, Line 27 (and references hereafter): I think you mean Solomon et al. (2011) (which discusses aerosols), rather than Solomon et al. (2010) (which discusses stratospheric water vapor).

Thanks for the corrections.

We have checked that carefully and corrected the errors.

Page 22122, Line 10: 2100 contradicts 2099 used in Table 1.

Thanks for the corrections.

We have checked the whole paper carefully and corrected the mistakes.

Page 22124, Equation 3: Please rewrite. It has two division signs.

We have rewritten this equation.

Page 22126, Line 11: Has should be have.

Corrected.

Page 22127, Line 25: Should be “Sect. 2.3”

Corrected.

Page 22128, Line 24: Text says “insignificant”, but figure shows shading.

Corrected.

Page 22131, Line 10: Change “errors” to “arrows”

Corrected.

Page 22131, Line 19: Change “divergence” to “convergence”

Corrected.

Page 22131, Line 27: 12-16 km is in the mid-latitude lower stratosphere.

Corrected.

Section 4.1: The term upwelling (instead of upward vertical wave propagation) is mistakenly used in this section several times. Please correct.

Thanks for this comment.

We have reconstructed the figure and also corrected the description in the revised manuscript.

Page 22133, Line 9 (and elsewhere): Transit branch? Do you mean shallow, or perhaps, transition branch?

Thanks for this comment.

We have checked that carefully, and used “lower branch” here, since the region (below 100 hPa) is actually lower than the transition branch (100-70 hPa, defined by Lin and Fu, 2013).

Thank you very much for these technical comments. We have addressed all of them and highlighted the respective changes in the text.

References

Lin, P. and Fu, Q.: Changes in various branches of the Brewer–Dobson circulation from an ensemble of chemistry climate models, J. Geophys. Res., 118, 73–84, doi:10.1029/2012JD018813, 2013.

Anonymous Referee #2

Received and published: 21 September 2014

This paper focuses on a problem of interest to the scientific community: understanding the trends in the temperatures around the tropopause. Nevertheless, this paper falls far short of what is necessary for publications.

The basic approach of this paper is to take observations over a 10-year period and compare those to a climate model simulation in order to determine the factors that control the trend. There are major problems, however, with this approach that lead me to conclude that the results of this paper are simply not reliable.

1. Model does not reproduce the trend: it is not stated as explicitly as it should be, but in Sections 4 and 5 it is revealed that the model produces a TTL trend that is much smaller than the 10 years of observations. In fact, in the final summary, more than half of the observed trend is attributed to errors in the model's vertical resolution. Given such a large difference between the model and the observations, how can you trust that the model tells us anything about the observed trend?

Response:

We agree with the reviewer that the simulated TTL variability trend is smaller than in observations and that it is highly sensitive to the period used for the analysis (see also our comment in reply to reviewer 1 and the two additional supplementary material). The smaller modeled TTL variability does not indicate a failure of the model. We show that with finer vertical resolution, the model captures the observed TTL variability better (Wang et al., 2013). However, the simulations using a fully interactive chemistry and ocean module are so demanding in computer time and real time resources, that it was impossible to complete all simulations with the high vertical resolution version of the model. The idea of this paper is instead to use the low vertical resolution simulations, which are standard in most CCMs and GCMs, and estimate the general contributions of different natural and anthropogenic variability factors, such as solar GHGs and aerosol, to the TTL variability.

For this estimate we use a fully coupled atmosphere-ocean model which represents better the processes needed to study TTL variability. It is well known that using SSTs as lower boundary conditions only without interactive feedback between the ocean and the atmosphere may have problems in reproducing a "correct" atmospheric variability. Here we investigate the influences of different factors on TTL temperatures in a fully-coupled way, and at the same time highlight the importance of finer model vertical resolution while using SSTs as a climate forcing.

Changes in manuscript:

We have extended our analysis method and hope that we explain better now what our main findings are.

It is an interesting conclusion that vertical resolution has such an important effect, but it calls into question all of the other conclusions about attribution of the observed trend. One could write an entire paper about the effect of vertical resolution, but the paper would be quite different from this paper (it would, for example, not contain Section 3).

Thank you very much for this suggestion. As a next step, we are indeed planning to investigate the effects of finer vertical resolution in more detail.

2. 10-year trends are unreliable: Anyone who has done any kind of atmospheric data analysis knows that looking at a 10-year trend is fraught with danger. In particular, a few outlier months can really torque the trend, so the model must simulate the yearto- year variability really well. The big worry here is that there is short-term variability in the atmosphere that is not accurately captured by the model. The model does use observed SST, but I don't see any reason to expect that this therefore includes all the short-term variability.

Thank you very much for your comments. We agree that a 10-year dataset is relatively short to investigate robust trends. We have added supplementary figures as well as a discussion about uncertainties of the observed trends and the sensitivity to the time period employed.

Since the observed dataset is so short, we used CESM-WACCM to perform longer simulations (145 years each) and calculate more statistically reliable signals from these longer model records. We use composite analysis to avoid the problem of different interanual variability in the coupled model runs and observations. For our WACCM atmosphere-only simulations, especially for the W_Aerosol run, we have observed SSTs, almost real GHGs and ODSs, observed solar irradiance, nudged QBO based on observed winds and observed stratospheric aerosols to capture the recent variability. And we also did 3 ensembles for the WACCM atmosphere-only simulations except for the W_Aerosol run to reduce the uncertainties by short simulations.

As an example, it is well-known that Brewer-Dobson variability has a big impact on TTL temperatures. Is the model getting the right BD circulation variability, with the right phase? Lack of correctly simulating this variability could be one of the reasons that the model does not reproduce the observed TTL trends. Note that using a much longer time series would avoid these problems.

We agree, the B-D circulation is important for TTL temperatures. However, as mentioned in the introduction, the strengthening or weakening of the B-D circulation is still an open question. One of our conclusions is that the finer vertical resolution may be important for representing the variability in the B-D circulation, which is consistent with previous work (Bunzel and Schmidt (2013)).

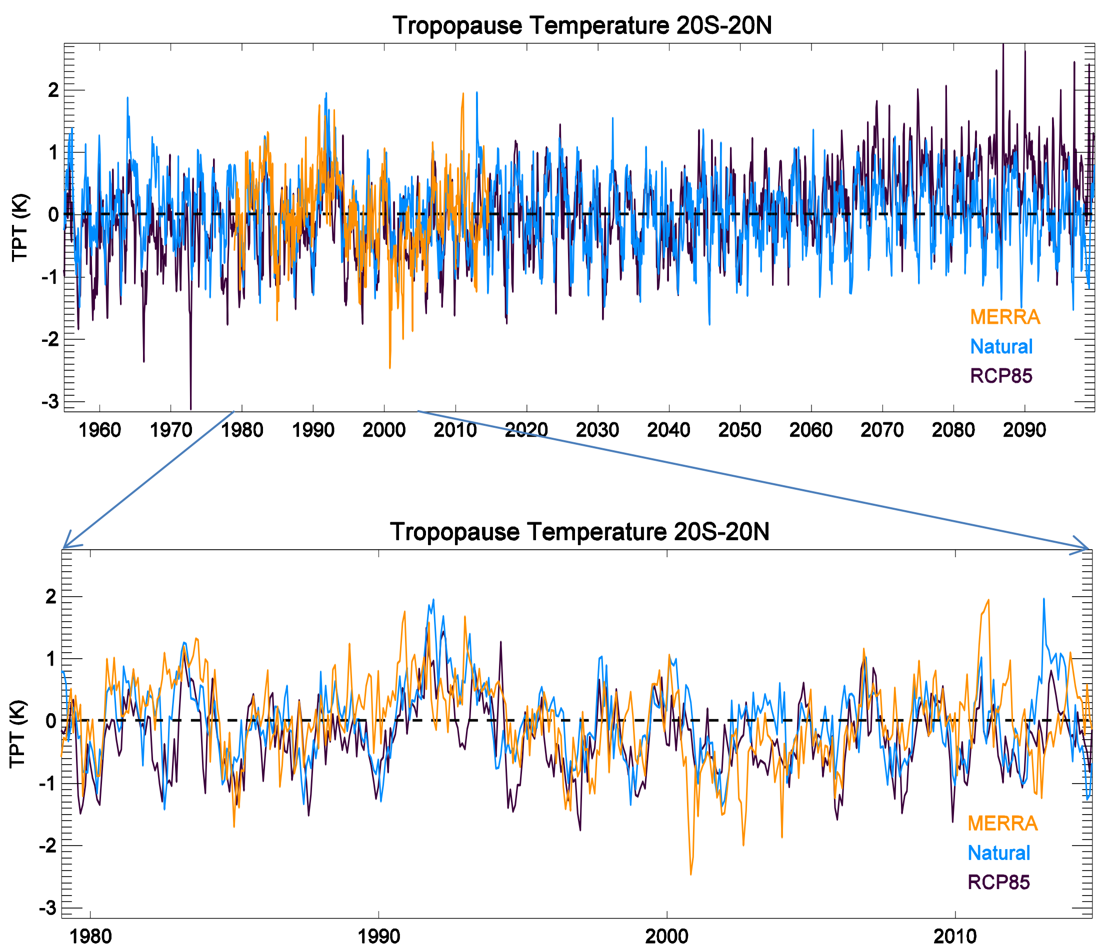
3. no autocorrelation in error estimates: Based on the discussion of error bars, it does not appear that the authors have taken autocorrelation of the time series into account in calculating error bars on the trend. All of the time series considered here are autocorrelated – meaning that a month is more likely to be high if the previous month was high – and this means that there are fewer independent samples in the time series than there are months. As a result, I suspect that the error bars will be larger than those presented here, which will reduce the statistical significance of the paper's conclusions. If the authors choose to revise this paper, they must recalculate the error bars to incorporate autocorrelation.

Thank you very much for raising the autocorrelation issues (also requested from reviewer 1). We have reconsidered our method how to get to statistically significant tests with autocorrelation considerations. See details of the methods and results in the revised manuscript.

4. paper's conclusions make no sense: Finally, the conclusions of the paper make no sense. It is well known, for example, that tropopause-level temperatures increase over the 21st century in climate models. It seems virtually certain that this is due to some combination of increasing surface temperature and increasing greenhouse gas concentrations (after all, what else could it be?). However, this paper concludes the opposite: that warming SST and increasing greenhouse gases COOLS the TTL. Given that this goes against every other analysis of models that I've seen, I strongly suspect that problems in the methodology (discussed above) are the cause of this highly curious conclusion. If the author's revise this paper, they have to more directly explain these results.

We apologize to have contributed to confusion on this point. We couldn't find any references indicating a well known long-term increase of tropopause temperatures over the 21st century. Neither of the CCMVal and CMIP5 models can reproduce the cooling in tropical tropopause temperatures during some decades shown in observational and reanalysis data. The model predictions of the long-term trend are also highly uncertain and have large spreads (CCMVal report chapter 7, Kim et al., 2013 JGR).

The temperature structure and variability around the tropical tropopause, i.e. in the tropical tropopause layer (TTL), is very complex, and can be influenced by a number of different factors. This includes natural and anthropogenic factors as discussed in this paper but also other factors such as lower stratospheric ozone and water vapor. The dynamics and thermal characteristics of the TTL are important in determining stratospheric composition, affecting radiative and dynamical properties at both stratospheric and tropospheric levels. In particular, the coldest level in the TTL, the cold-point tropopause (CPT), is known to play a crucial role in stratosphere-troposphere exchange (Holton et al ., 1995). The CPT temperature largely determines the concentration of water vapor in the lower stratosphere, which serves as a key radiative constituent for surface climate (e.g., Solomon et al ., 2010).



The Figure above shows the time series of tropical tropopause temperature anomalies from the past to the future in MERRA reanalysis data and our Natural and RCP85 runs. Obviously, the tropical tropopause temperature has large interanual fluctuations in both MERRA data and the two CESM runs. The CESM model simulates similar interanual (bottom), decadel to multidecadal (top) variability in the tropical tropopause temperature compared to MERRA data over the time period 1979 through 2014. Even with a very strong GHG scenario RCP8.5 (black line), the tropical tropopause temperature only shows a slight increase after 2050. Understanding the contributing factors to this decadal tropical tropopause temperature variability is the goal of our paper. We come to the conclusion that this decadal variability seems to be related to internal climate variability.

We have added this Figure to the revised manuscript.

I wish I could be more encouraging, but in the end I am not convinced that the results of this paper are correct. There are too many methodological errors and logical flaws in it for the results to be considered reliable.

We hope to have addressed all the concerns of you with our detailed reply and the additional analysis and changes to the text.

References

Bunzel, F. and Schmidt, H.: The Brewer–Dobson circulation in a changing climate: impact of the model configuration, J. Atmos. Sci., 70, 1437–1455, doi:10.1175/JAS-D-12-0215.1, 2013.

Kim, J., K. M. Grise, and S.-W. Son (2013), Thermal characteristics of the cold-point tropopause region in CMIP5 models, J. Geophys. Res. Atmos., 118, 8827–8841, doi:10.1002/jgrd.50649.

Anonymous Referee #3

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Review of "Quantifying contributions to the recent temperature variability in the tropical tropopause layer" by Wang et al.

This paper investigates natural and anthropogenic contributions to the decadal variability of the tropical tropopause layer (TTL) temperature. Using a series of sensitivity experiments with NCAR's CESM model, the authors quantify the impacts of solar cycle, SST, QBO, stratospheric aerosol and greenhouse gas increase on the observed TTL warming in the 2001-2011 period. They find that the recent TTL warming is mainly caused by tropical SST decrease and QBO amplitude increase. This paper also highlights the importance of using high vertical resolution in order to correctly simulate TTL decadal variability.

Results presented in this paper are important to understand decadal variability in the TTL. I have some comments, especially on the design of model experiments, to improve the manuscript. I recommend publication after my comments are addressed.

Comments:

What is the benefit of using the fully-coupled CESM-WACCM instead of WACCM? It appears to me that the method used with the stand-alone WACCM runs (section 2.2) is more straightforward. All the runs listed in table 1 can be conducted with the standalone WACCM.

Response:

The atmosphere is the primary source of internal variability to the atmosphere-ocean system, especially in midlatitudes. The coupling between the atmosphere and the ocean can decrease the energy flux between the atmosphere and the ocean. Using SSTs as the lower boundary forcing for an atmospheric model may not generally lead to a correct simulation of low-frequency atmospheric thermal variance (Barsugli and Battisti, 1998 JAS; Wang et al., 2005 GRL).

Page 22122, How the nudged QBO is done in CESM? I know there are references, but it would be better to have a brief description in the text.

Response:

We appreciate the proposal to extend the QBO nudging procedure description in the paper.

Changes in manuscript:

We have adapted the text as:

The QBO is nudged by relaxing the modeled tropical winds to observations, which using a Gaussian weighting function decaying latitudinally from the equator with a half width of 10°. Full vertical relaxation extends from 86 to 64 hPa, which is half that strong in one model level below and above this range and zero for all other levels (see details in Matthes et al., 2010; Hansen et al., 2013).

Page 22125, last paragraph, Why don't just repeat the 2001-2011 solar cycle in the nature run? Then there would be more samples to compare with the SolarMean run.

It is true that by repeating only one solar cycle we could generate more samples and we agree with the reviewer that this would be a very valuable exercise. Unfortunately, the simulations with interactive chemistry and interactive ocean are computationally very expensive and we can only do a limited amount of experiments. The Natural run was designed to be a control experiment for other studies we are performing some of which are hindcasts of the recent past and require a realistic solar cycle forcing.

Page 22129, section 3.4, Why increased QBO leads to warming in the TTL? Maybe it is related to the weakening of upwelling in the TTL. If this is true, I suggest the authors compare differences in upwelling in the control and NOQBO runs.

Thank you for this comment. As indicated by Kawatani and Hamilton (2013), the QBO and tropical upwelling are closely related to each other. We have compared the tropical upwelling between the control and the NOQBO run, and the differences in tropical upwelling are highly correlated to the QBO time series in the control run. We also regressed temperature differences between these two runs onto both, the time series of the QBO in the control run, and the differences in tropical upwelling. The regressed temperature differences are very similar to each other. The results above indicate that the QBO has indeed an impact on the TTL temperatures very likely by modifying the tropical upwelling.

We have added a brief discussion in the revised manuscript.

Section 4, It's difficult to follow the discussions. Changes in the residual circulation shown in Figures 10c and 10d are complicated and not easy to explain. I suggest the authors using simple diagnostics, e.g., mean upwelling in the TTL and lower stratosphere, to illustrate differences in the BDC in the high and low resolutions runs.

We appologize for the inconvenience with this Figure. The other reviewers have also suggested to simplify the figures and to sharp the discussion.

We have combined Figures 9 and 10 and restructured the figure to be better readable.

Figure 10b shows a strong cooling trend in the Antarctic lower stratosphere in the high resolution run? What causes this cooling? Is it related to the Antarctic ozone hole? A more general question is how stratosphere ozone depletion might affect TTL temperature variability.

Thank you very much for this suggestion. We have checked the ozone differences in both runs W_L103 and W_L66, and the ozone shows indeed a strong decrease in the Antarctic . Ozone depletion therefore seems to play some role to the TTL temperature variability. Since we changed the Figure, the cooling is no longer in the figure and we therefore prefer to discuss only the effects in the tropics and subtropics. We agree that an extension of the analysis to higher latitudes would be interesting, but this is beyond the scope of the current paper.

Summary, I think it would be helpful to add a simple figure summarizing the contribution of different factors to the TTL decadal variability.

Thank you very much for this proposal.

We have added a table to summarize the contribution of the different factors in the revised manuscript.

References

Joseph J. Barsugli and David S. Battisti, 1998: The Basic Effects of Atmosphere–Ocean Thermal Coupling on Midlatitude Variability. J. Atmos. Sci., 55, 477–493.

Kawatani, Y. and Hamilton, K.: Weakened stratospheric quasibiennial oscillation driven by increased tropical mean upwelling, Nature, 497, 478–481, doi:10.1038/nature12140, 2013.

Wang, B., Q. Ding, X. Fu, I.-S. Kang, K. Jin, J. Shukla, and F. Doblas-Reyes (2005), Fundamental challenge in simulation and prediction of summer monsoon rainfall, Geophys. Res. Lett., 32, L15711, doi:10.1029/2005GL022734.

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Quantifying contributions to the recent temperature variability in the tropical tropopause layer

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Abstract

The recently observed variability in the tropical tropopause layer (TTL), which features an unexpected warming of ~~1.1~~1.0 K over the past decade (2001–2011), is investigated with a number of sensitivity experiments from simulations with NCAR's CESM-WACCM ~~chemistry climate~~ chemistry-climate model. The experiments have been designed to specifically quantify the contributions from natural as well as anthropogenic factors, such as solar variability (Solar), sea surface temperatures (SSTs), the Quasi-Biennial Oscillation (QBO), stratospheric aerosols (Aerosol), greenhouse gases (GHGs), as well as the dependence on the vertical resolution in the model. The results show that, in the TTL: a cooling in tropical SSTs leads to a weakening of tropical upwelling around the tropical tropopause and hence relative downwelling and adiabatic warming of ~~0.3~~0.4 K decade^{-1}; an increased QBO amplitude results in a ~~0.3~~0.5 K decade^{-1} warming; increasing aerosols in the lower stratosphere lead to a 0.4 K decade^{-1} warming; a prolonged solar minimum and increased GHGs contribute about ~~0.2~~ 0.3 and 0.1 K decade^{-1} to a cooling, respectively. Two simulations with different vertical resolution show that the vertical resolution can strongly influence the response of the TTL temperature to changes such as SSTs. With higher vertical resolution, an extra ~~0.6~~0.8 K decade^{-1} warming can be simulated through the last decade, compared with results from the ~~“standard”~~ "standard" low vertical resolution simulation. Considering all the factors mentioned above, we compute a net ~~1.3~~1.7 K decade^{-1} warming, which is in ~~very~~ good agreement with the observed ~~1.1~~1.0 K decade^{-1} warming over the past decade in the TTL. The model results indicate that the recent warming in the TTL is mainly due to internal variability, i.e. the QBO and tropical SSTs.

1 Introduction

The ~~tropical tropopause layer (TTL)~~ TTL is the transition layer from the upper troposphere to the lower stratosphere in the tropics, within which the air has distinct properties of both the troposphere and the stratosphere. The vertical range of the TTL depends on how it is defined, i.e., it can be a shallower layer between 14–18.5 km (Fueglistaler et al., 2009) or a deeper layer of about 12–19 km (Gettelman and Forster, 2002; SPARC-CCMVal, 2010, chapter 7). As a key region for the stratosphere-troposphere coupling, the TTL acts like a “gate” for air ~~masses entering~~ entering into the stratosphere from the tropical troposphere. The ~~coupling between dynamics, radiation and chemistry is especially strong in the TTL since it is the source region for trace gases entering into the stratosphere. The~~ temperature in the TTL is determined by the combined influences of latent heat release, thermally as well as ~~mechanically~~ dynamically driven vertical motion, and radiative cooling (Gettelman and Forster, 2002; Fueglistaler et al., 2009; Grise and Thompson, 2013). The thermal structure, static stability and zonal winds in the TTL affect the two-way interaction between ~~troposphere and~~ the troposphere and the stratosphere (Flury et al., 2013; Simpson et al., 2009) as well as the surface climate, since the relative minimum temperature (usually known as the cold point tropopause, CPT) subsequently influences the radiation and water vapor budget (Andrews, 2010). The TTL reacts particularly sensitively to anthropogenically induced radiative, chemical and dynamical forcings of the climate system, and hence is a useful indicator ~~of~~ for climate change (Fueglistaler et al., 2009).

Over the past decade, a remarkable warming has been captured by Global Positioning System Radio Occultation (GPS-RO) data in the TTL region ~~see also an extension in Fig. 2Schmidt2010, wang2013recent.~~ (Schmidt et al., 2010; Wang et al., 2013). This might indicate ~~for~~ a climate change signal, with possible important impacts on stratospheric climate, ~~i.e.~~e.g., the tropical tropopause temperature dominates the water vapor entering the stratosphere (Dessler et al., 2013; Solomon et al., 2010; Gettelman et al., 2009; Randel and Jensen, 2013). This recent warming of the TTL is different to previous reported temperature evolutions over the last decades. So far a long-term cooling has been reported from the 1970s to 2000, although there are large differences between different datasets (Wang et al., 2012). The recent warming is therefore of great interest to study the exact reason for this warming in the TTL which might indicate a climate shift. An interesting question is also whether this warming will continue or change in sign in the future, and how well climate models can reproduce such a strong warming over one decade or longer time periods.

Based on model simulations, Wang et al. (2013) suggested that the warming around the tropical tropopause could be a result of a weaker tropical upwelling, which implies a weakening of the Brewer–Dobson circulation (BDC). However, the strengthening or weakening of the BDC is still under debate (Butchart, 2014, and references therein). Results from observations indicate that the BDC may have slightly decelerated (Engel et al., 2009; Stiller et al., 2012), while estimates from a number of ~~CCMs~~ Chemistry-Climate Models (CCMs) show in contrast a strengthening of the BDC ~~(Butchart et al., 2010; Li et al., 2008)~~(Butchart et al., 2010; Li et al., 2008; Butchart, 2014). The reason of the discrepancy between observed and modeled BDC changes, as well as the mechanism ~~for changes in~~ of the BDC in response to climate change, is still under discussion (Oberländer et al., 2013; Shepherd and McLandress, 2011). The trends in the BDC may be different in ~~the~~ different branches of the BDC (Lin and Fu, 2013; Oberländer et al., 2013). Bunzel and Schmidt (2013) show that the model configuration, i.e. the vertical resolution and the vertical extent of the model, can also impact trends in the BDC.

There are a number of other natural and anthropogenic factors besides the BDC, which influence the radiative, chemical and dynamical processes in the TTL. One prominent candidate for natural variability is the sun, which provides the energy source for the climate system. The 11 year solar cycle is the most prominent natural variation on the decadal time scale (Gray et al., 2010). Solar variability influences the temperature through direct radiative effects and indirectly through radiative effects on ozone as well as indirect dynamical effects. The maximum response in temperature occurs in the equatorial upper stratosphere during solar maximum conditions, and a distinct secondary temperature maximum can be found in the equatorial lower stratosphere around 100 hPa (SPARC-CCMVal, 2010; Gray et al., 2010). SSTs also influence the TTL by ~~influencing~~ affecting the dynamical conditions and subsequently the propagation of planetary waves and hence the circulation. Increasing tropical SSTs can enhance the BDC, which in turn cools the tropical lower stratosphere through enhanced upwelling (Grise and Thompson, 2012, 2013; Oberländer et al., 2013). The QBO is the dominant mode of variability throughout the equatorial stratosphere, and has important impacts on the temperature structure as well as the distribution of chemical constituents like water vapor, methane and ozone (Baldwin et al., 2001). ~~Stratospheric aerosols warm the lower stratosphere by reflecting and scattering incident solar radiation back to space or by absorbing outgoing long-wave radiation, with a maximum aerosol heating~~ Stratospheric aerosols absorb outgoing long-wave radiation and lead to additional heating in the lower stratosphere, which is maximized around 20 km ~~solomon2010contributions~~(Solomon et al., 2011; SPARC-CCMVal, 2010, chapter 8).

While GHGs warm the troposphere, they cool the stratosphere at the same time by releasing more radiation into space. Warming of the troposphere and cooling of the stratosphere affect the temperature in the TTL directly, and also indirectly, by changing chemical trace gas distributions and wave activities (SPARC-CCMVal, 2010).

~~A sufficiently~~ In climate models, a sufficient high vertical resolution is important in order for models to correctly represent dynamical process, such as wave propagation into the stratosphere and wave-mean flow interactions. It is a prominent factor for a climate model to generate a self-consistent QBO (Baldwin et al., 2001; Bunzel and Schmidt, 2013)(Richter et al., 2014). Meanwhile, vertical resolution is essential ~~to~~ for a proper representation of the thermal structure in the model, e.g. models with coarse vertical resolution can not simulate the tropopause inversion layer (TIL, a narrow band of temperature inversion above the tropopause associated with a region of enhanced static stability) well (Wang et al., 2013; SPARC-CCMVal, 2010, chapter 7). Coarse vertical resolution is also still a problem for analysing the effects of El-Niño Southern Oscillation (ENSO) and the QBO onto the tropical tropopause (Zhou et al., 2001; SPARC-CCMVal, 2010, chapter 7).

~~Here~~ In this study we use a series of simulations with NCAR's Community Earth System Model (CESM) model (Marsh et al., 2013), to quantify the contributions of the above discussed ~~factors-Solar~~factors – Solar, SSTs, QBO, Aerosol and ~~GHGs-to~~ GHGs – to the recently observed variability in the TTL.

The details of the observed data, the model and numerical experiments, as well as a description of our methods are given in Sect. 2. The observed temperature variability in the

TTL and the contributions of various factors to the recent TTL variability are addressed in Sect. 3. Section 4 focuses on the importance of the vertical resolution in one climate model. A summary and discussion are presented in Sect. 5.

2 Model simulations and method description

2.1 Fully-coupled CESM-WACCM simulations

The model used here is NCAR's Community Earth System Model (CESM)~~model~~, version 1.0. CESM is a fully coupled model system, including an interactive ocean (POP2), land (CLM4), sea ice (CICE) and atmosphere (CAM/WACCM) component (Marsh et al., 2013). As the atmospheric component we use the Whole Atmosphere Community Climate Model (WACCM), version 4. WACCM4 is a ~~chemistry climate~~ chemistry–climate model (CCM), with detailed middle atmospheric chemistry and a finite volume dynamical core, extending from the surface to about 140 km (Marsh et al., 2013). The standard version has 66 (W_L66) vertical levels, which means about 1 km vertical resolution in the TTL and in the lower stratosphere. All ~~the~~ simulations use a horizontal resolution of $1.9^{\circ} \times 2.5^{\circ}$ (latitude $\times$ longitude) for the atmosphere and approximately 1 degree for the ocean.

Table 1 gives an overview of all coupled CESM simulations. A control run was performed from 1955 to ~~2100~~ 2099 (Natural run hereafter), with all natural forcings including spectrally resolved solar variability (Lean et al., 2005), a fully coupled ocean, volcanic aerosols ~~according to CCMVal2 recommendations~~ (Morgenstern et al., 2010) ~~and nudged QBO~~ following the SPARC CCMVal REF-B2 scenario recommendations (see details in SPARC-CCMVal, 2010) and a nudged QBO. The QBO is nudged by relaxing the modeled tropical winds to observations between 22° S and N, using a Gaussian weighting function with a half width of 10° decaying latitudinally from the equator. Full vertical relaxation extends from 86 to 64 hPa, which is half the strength of the level below and above this range and zero for all other levels (see details in Matthes et al., 2010; Hansen et al., 2013). ~~The solar variability and volcanic aerosols used here follow the SPARC CCMVal REF-B2~~

~~scenario (see details in~~ SPARC-CCMVal, 2010~~).~~ The QBO forcing time series in CESM is determined from the observed climatology of 1953-2004 via filtered spectral decomposition of that climatology. This gives a set of Fourier coefficients that can be expanded for any day and year in the past and the future. Anthropogenic forcings like GHGs and ozone depleting substances (ODSs) are set to constant 1960s conditions. Using the Natural run as a reference, a series of four sensitivity experiments were performed by systematically switching on or off several factors. The SolarMean run uses constant solar cycle values averaged over ~~past~~ the past 4 observed solar cycles; the FixedSST run uses monthly varying climatological SSTs calculated from the Natural run, and therefore neglects variability from varying SSTs such as ENSO; in the NOQBO run the QBO nudging has been switched off which means weak zonal mean easterly winds develop in the tropical stratosphere. An additional simulation RCP85, uses the same forcings as the Natural run, but in addition includes increases in anthropogenic GHGs and ODSs forcings. These forcings are based on observations from 1955 to 2005, after which they follow the Representative Concentration Pathways (RCPs) RCP8.5 scenario (Meinshausen et al., 2011).

2.2 WACCM atmospheric stand-alone simulations

Instead of using the fully coupled CESM-WACCM version, WACCM can be integrated in an atmospheric stand-alone configuration, with prescribed SSTs and sea ice. Beside the standard version with 66 vertical levels (W_L66), we have also performed simulations with a finer vertical resolution, with 103 vertical levels and about 300 m vertical resolution in the TTL and lower stratosphere (W_L103) (Gettelman and Birner, 2007; Wang et al., 2013).

With the stand-alone atmospheric version an ensemble of three experiments was performed over the recent decade 2001–2010 with both WACCM versions (W_L66, W_L103) (see Table 2). Observed SSTs and spectrally resolved solar fluxes were used to produce the most realistic simulations of atmospheric variability over the past decade (2001–2010). The QBO is nudged using the same method as in the fully-coupled runs discussed above. GHGs and ODSs are based on observations for the first 5 years (2001–2005) and then follow the IPCC RCP4.5 scenario for the next 5 years (2005–2010), since no observed data were available when the simulations were started. Atmospheric aerosols were relatively constant between 2001 and 2010 since no strong volcanic eruptions occurred, and are the same as in the CESM-WACCM runs described above. All the forcings considered in this study are available from the CESM model input data repository (https://svn-ccsm-inputdata.cgd.ucar.edu/trunk/inputdata/). ~~The results presented in the following are ensemble averages for each of the two WACCM vertical resolution versions.~~ An additional run (W_Aerosol) was performed using the W_L103 version with observed, more realistic stratospheric aerosol forcings from the Chemistry-Climate Model Initiative (CCMI, http://www.met.reading.ac.uk/ccmi/) project.

2.3 Linear trend calculation

A standard least square regression is used to calculate the linear trend. For example, using time (t) as a single predictor, the predicted temperature (T) can be expressed as:

$$T_{\text{est}} = a + bt, \tag{1}$$

where the subscript "est" indicates that this is an estimate of T, and "b" represents the linear trend. The residuals are defined by the differences between the actual and the estimated temperature

$$e = T - T_{\text{est}}. \tag{2}$$

The "best-fit" is defined by the line that minimizes the sum of the squares of the residuals. The seasonal cycle was removed from the temperature time series before doing the regression.

The standard error (SE) is used to estimate the uncertainty of the estimated trend, which is defined by:

$$(\text{SE})^2 = \left[\sum e^2\right] / ((n-2)\cancel{/}\left[\sum\left((t-\bar{t})^2\right)\right]), \tag{3}$$

where n is the sample size, and $\bar{t}$ is the mean value. The smaller this standard error is, the more certain is the trend.

The formula given above is, however, only correct if the individual data points are statistically independent. Considering the effect of autocorrelation, the sample size n will be reduced and an "effective sample size" can be determined by:

$$n_{\text{eff}} = n(1-r)/(1+r), \tag{4}$$

where r is the lag-1 autocorrelation coefficient.

If the trend is larger than two times the standard error, the linear regression is ~~statically~~ statistically significant at the 95 % confidence level. A brief description of the least square regression, the uncertainty of the trend, as well as the significance testing can be found in Wigley (2006).

2.4 Composite method

To estimate the contribution of the different factors to the observed temperature variability in the TTL, the composite for each factor is computed following three steps: (1) calculation of the linear trend for the respective factor over the past decade (2001–2011) from the observed, deseasonalized time series, (2) selection of time periods in the reference run, which are similar to the observed trends for the respective factor (the method of selection and the number of similar time periods depends on the factor, see below), and (3) ~~calculation of the linear temperature trends for each selected time period, followed by averaging of all trend periods together to obtain a composite trend. The composite trend is calculated for both each reference run and the run to which it is compared. Both runs have the same configuration and forcings, except for the long-term variability of the respective factor. The contribution of this factor is then quantified by the difference between the reference run and the run without this factor.~~ calculation of the composite trend in temperature from all selected time periods by multiple linear regression.

~~The difference of the trend between two runs is simply the difference time series, i.e. subtracting the time series of the reference run from the run to which it is compared. The uncertainty and the significance of the trend difference between two runs are then estimated by the standard error of the regression in the difference time series and the comparison between the trend and the standard error.~~ With multiple (np) selected periods, the least square regression can be expressed as:

$$T_{\text{est}} = a + bt + c_1 t_1 + \ldots + c_{\text{np}} t_{\text{np}}, \tag{5}$$

where t is the time in all selected periods and b represents the composite trend, $c_1 \ldots c_{\text{np}}$ are used to estimate the "jumps" between different periods, $t_1 \ldots t_{\text{np}}$ are set to 1 for selected periods 1 $\ldots np$, and equal to 0 for other periods.

~~The uncertainty of the composite trend is estimated by the mean of the standard errors in different selected time periods. The significance of the composite trend is then tested by a comparison between the composite trend and the mean of the standard error.~~

~~Special attention is given to the region 20~~As described above, the "effective sample size" (n_{eff}) is determined by the total sample size n and the autocorrelation coefficient r. The standard errors (SE) can be estimated from equation (3) with the degrees of freedom $n_{\text{eff}} - np - 2$. For the estimated trend b, the test statistics

$$\tag{6}$$

$$SUBSCRIPTNBtS--20Nlatitudeand16--18.5height, whichisalmostexactlytheTTLreg$$

$$\tag{7}$$

has the Student's t-distribution with $n_{\text{eff}} - np - 2$ degrees of freedom.

The composite trend is then calculated for each of the two runs (e.g. Natural and SolarMean) to be compared as well as for their differences. Both runs have the same configuration and forcings, except for the long-term variability of the respective factor (e.g. Solar).

The contribution of this factor is then quantified by the trend differences between the two runs.

Special attention is given to the region 20° S–20° N latitude and 16–20 km height, which is mainly the observed warming area in the TTL (see below). Hereafter, we use the average trend over this area to discuss the exact contribution of every factor to the temperature trend in the TTL.

2.5 Forcings in observations and model simulations

Figure 1 shows the time series of both natural and anthropogenic forcings over past and future decades in the observations (black) and in the Natural model experiment (blue). Periods with a similar trend as the recent decade (2001–2011) are shown with straight lines.

Observations of the solar variability show that the total solar irradiance (TSI) exhibits a clear 11 year solar cycle (SC) variation of about 1 W m^{-2} between sunspot minimum (S_{min}) and sunspot maximum (S_{max}) in the past (Gray et al., 2010), with a delayed and smaller amplitude return to maximum conditions in the recent decade (Fig. 1a). The future projection is a repetition of the last four observed solar cycles. Similar periods of decreasing TSI can be found in the periods 1958–1968, 2001–2011, 2045–2055, and 2089–2099. A composite trend is then constructed by ~~averaging over~~ applying multiple linear regression to all four selected periods in the Natural as well as the SolarMean experiments ~~.~~ (Eq. 5). By comparing the trends in these two runs with and without solar cycle variations, the effect of solar variability on the temperature trend in the TTL can be estimated.

Figure 1b shows the variability of tropical (20° S–20° N) SSTs for the ~~recent past~~ last five decades from observations (Hadley Center Updates and supplementary information available from http://www.metoffice.gov.uk/hadobs/hadisst, black lines) and up to ~~2100~~ 2099 from the Natural coupled CESM-WACCM model experiment (blue ~~lines~~line). Both the observed and simulated tropical SSTs show a statistically significant (over 95 %) decrease from 2001 to 2011. A similar decrease in tropical SSTs can be found during the periods 1956–1968, 1980–1991, 2001–2014, 2028–2043 and 2047–2057. Periods longer than 10 years ~~each has~~ have been selected from the filtered tropical SST time series. Filtering has been performed twice with a Butterworth low-pass filter (longer than 30 years). ~~Afterwards a composite trend was constructed by averaging over the five selected time periods.~~ Note that there is also a strong drop in SSTs around 1992 in the model, which does not occur in the observations. This might be caused by an overestimated response to the Pinatubo eruption in CESM-WACCM (Marsh et al., 2013; Meehl et al., 2012). By comparing the Natural run, where SSTs are calculated explicitly, and the FixedSST run where SSTs are climatologically prescribed, the effect of interactively calculated SSTs can be determined.

Instead of a decrease in tropical SSTs, the QBO amplitude shows an increase during the selected two periods (Fig. 1c). The observed QBO amplitude has been calculated from the absolute values of deseasonalized monthly mean anomalies of the zonal mean zonal wind at 70 hPa (from the FU Berlin: http://www.geo.fu-berlin.de/en/met/ag/strat/produkte/qbo/index.html); in our model simulations it is computed where the QBO has been nudged. The time periods of increasing QBO amplitude are selected by the same as the procedure as for the tropical SSTs. By comparing the Natural and the NOQBO experiments, the effect of a (nudged) QBO on the temperature trends in the TTL can be estimated.

As shown in Fig. 1d, GHGs show a steady increase after 2001. The increasing rate of global CO_2 release from 2001 to 2011 is close to the RCP8.5 scenario, which we used in our RCP85 run. By comparing the experiments with (~~RCP8.5~~RCP85) and without (Natural) GHG increases, the GHG effect on the observed temperature trend can be estimated. To avoid possible uncertainties due to the short time series, a longer time period from 2001 through 2050 will be used to compare the Natural and the RCP85 run.

Similar to the GHGs, observed stratospheric aerosols (aerosol optical depth (AOD)) have been steadily increasing since 2001 (Solomon et al., 2010) (Solomon et al., 2011) in the lower stratosphere (18–32 km) (Fig. 1e). This increase in stratospheric aerosol loading is attributed to a number of small volcanic eruptions and anthropogenically released aerosols transported into the stratosphere during the ~~Asia~~ Asian Monsoon (Bourassa et al., 2012; Neely et al., 2013). An aerosol data set has been constructed for the CCMI project (ftp://iacftp.ethz.ch/pub_read/luo/ccmi/) and is similar to the data described by Solomon et al. (2010)Solomon et al. (2011). The comparison of the two experiments with different AOD data sets will shed light on the stratospheric aerosol contribution to the observed temperature trend.

All natural and anthropogenic forcings will be discussed with respect to their contribution to the temperature variability in TTL in the following section.

3 Quantification of observed temperature variability

3.1 Observed temperature variability in the TTL

Figure 2 shows the latitude-height section of the linear temperature trend for the period 2001–2011 estimated from GPS-RO observations (see details of the GPS-RO data in Wang et al., 2013). A remarkable and statistically significant warming occurs around the TTL between about 20° south to north and from 15 to 20 km height. The warming in the TTL is ~~1.1 ± 0.2~~1.0 K decade^{-1} on average, with a maximum of about 1.8 K decade^{-1} directly at the tropical tropopause around 17 to 18 km. ~~The uncertainty is given by the standard error of the trend as described in Sect. 2.4.~~ This figure is an extension of earlier work by Schmidt et al. (2010) and Wang et al. (2013) and shows an unexpected warming, despite the steady increase in GHGs which imply a cooling of this region. Therefore it is interesting to study whether this warming is simply a phenomenon of the past decade and the result of internal atmospheric variability, or whether it will persist for longer and therefore modify trace gas transport from the troposphere into the stratosphere.

Please note that this decadal warming in the TTL may vary in magnitude if different end years are selected due to the relative short length of the time series. The warming is weaker if end years of 2012 or 2013 are chosen (seed also Figs. S1 and S2). In the following investigations, we keep the period from 2001 through 2011 to be most consistent with our WACCM simulations (2001–2010). We will explain the temperature variability within a time period of about one decade. This decadal variability may change sign from decade to decade if it is mainly caused by natural/internal variability. However, it is still very important to understand the reasons and mechanisms behind these internal variability modes as it might eventually enhance our decadal to multi-decadal predictive skills.

3.2 Contribution of solar variability

~~Figure~~Figures 3a and b ~~shows~~ show the composite temperature trends for periods with decreasing solar irradiance (1958–1968, 2001–2011, 2045–2055, and 2089–2099) for the Natural and SolarMean runs, respectively. The Natural run shows ~~an insignificant~~ a partially significant temperature increase from S_{max} to S_{min} of about ~~0.2~~0.1 K decade^{-1} ~~around the tropical tropopause, and a decrease at the tropical lower stratosphere,~~ on average around the TTL, while the SolarMean run shows on average a partially statistically significant temperature increase ~~between 0.2 and 0.6~~of 0.4 K decade^{-1} in the tropics and subtropics. Figure 3c shows the differences in temperature trends between the Natural ~~run~~ and the SolarMean experiments. Solar variability thus contributed to a cooling of ~~0.2 ± 0.2~~about 0.3 K decade^{-1} in the TTL ~~between 2001 and 2011. The relatively large uncertainty in the combined temperature trend indicates that there are large uncertainties in the estimated contribution due to solar variability~~ during periods with similar solar variability to the recent decade.

3.3 Contribution of tropical SSTs

Figure 4 shows the composite temperature trends for the periods with decreasing tropical SSTs (1956–1968, 1980–1991, 2001–2014, 2028–2043 and 2047–2057) for both the Natural and the FixedSST runs (Fig. 4a and b), as well as their differences (Fig. 4c). While the Natural experiment shows a partially statistically significant temperature increase of 0.4 K decade^{-1} on average ~~maximally 0.8~~(1.0 K decade^{-1} in maximum) in the TTL, the FixedSST experiment shows ~~only all~~ an insignificant (0–0.2 K decade^{-1}) ~~warming~~

~~during~~ temperature change during periods with similar SST variability to the recent decade. A decrease in tropical SSTs contributes therefore to a statistically significant warming of ~~0.3 ± 0.2~~0.4 K decade^{-1} on average (0.6 K decade^{-1} in maximum) in the TTL (Fig. 4c). ~~The uncertainty in the combined temperature trend reveals relative small uncertainties from the contribution of tropical SSTs.~~

3.4 Contribution of the QBO

~~Figure~~Figures 5a and b ~~shows the~~ show composite temperature trends, for periods with increasing QBO amplitudes (2003–2017 and 2054–2068) for the Natural and the NOQBO experiment, respectively. While the Natural experiment shows ~~a warming of 0.2~~an insignificant warming of 0.3 K decade^{-1} on average ~~maximally~~ (0.8 K decade^{-1} in maximum) in the TTL, the NOQBO run shows a slight but statistically significant cooling for the same period. The differences in Fig. 5c indicate that the increased QBO amplitude contributes to a warming of ~~0.3 ± 0.2~~0.5 K decade^{-1} on average (0.8 K decade^{-1} in maximum) in the TTL. ~~The uncertainty in the composite temperature trend indicates a relatively small uncertainty in the QBO contribution to TTL temperature.~~ Another effect of the QBO is the statistically significant cooling trend seen in the tropical middle stratosphere above 22 km in the Natural run. This QBO effect may help to explain the observed tropical cooling (see Fig. 2). Please note, however, that, ~~the~~ CESM1.0 ~~, which was~~ used for these simulations, ~~can not~~ cannot generate a self-consistent QBO and hence uses wind nudging, which might cause problems when estimating QBO effects on temperature variability in the tropical lower stratosphere (Marsh et al., 2013; Morgenstern et al., 2010).

3.5 Contribution of GHGs

The temperature trends from both the Natural and the RCP85 experiments between 2001 and ~~2011~~ 2050 are shown in ~~Fig~~Figs. 6a and b, respectively. As expected, increasing GHGs in the ~~RCP8.5 experiment tends~~ RCP85 experiment tend to cool the ~~TTL, whereas the contribution from the run with fixed GHG conditions is slightly positive~~lower and warm the upper troposphere, with a zero line slightly above the tropopause. Hence the effect of global warming is seen in Fig. 6c with a ~~clear~~ weak averaged cooling trend in the TTL of about ~~0.1 ± 0.2~~0.1 K decade^{-1}. ~~The uncertainty of the GHGs contribution is relatively large. Figure 6c shows that the GHGs do not affect the TTL temperatures much, but contribute instead to a cooling at upper levels in the lower stratosphere.~~ Further estimates were also calculated with different ending years for both the Natural and the RCP85 runs (not shown). The longer the chosen time period the clearer the patterns related to the GHG effect, i.e. a clear warming in the upper troposphere and a cooling in the lower stratosphere with the zero-line just above the tropopause.

3.6 Contribution of stratospheric aerosols

The temperature trends from the simulations with relative constant AOD values (W_L103) and with more realistic CCMI aerosols (W_Aerosol) are shown in ~~Fig~~Figs. 7a and b, respectively. A clearly stronger and more statistically significant warming pattern can be seen around the tropical tropopause in the W_Aerosol run as compared to the W_L103 run. The effect of increasing stratospheric aerosols is estimated to be 0.4 ~~± 0.2~~K decade^{-1} warming in the TTL (Fig. 7c). ~~The uncertainty of the contribution due to stratospheric aerosol is relatively small.~~ Please note that there may exist uncertainties for this result since we have only 10 years of simulations for the W_Aerosol run. With such a short time period, and after considering autocorrelation effects, the "effective sample size" will be strongly reduced and hence will cause all strong trends to be insignificant. The trends in the W_L103 run in contrast have been estimated by the ensemble mean of the three simulations and might not suffer from this problem.

~~4 Effect of the vertical resolution~~

To confirm the contributions of different factors from the composite analysis described above, a linear regression is additionally performed to estimate the contributions of Solar, SST, QBO and GHGs to the TTL temperature variability over the whole time series 1955–2099, from the Natural, SolarMean, FixSST, NOQBO and RCP85 runs. The time series of TTL temperature differences (comparative run minus control run, e.g., RCP85 - Natural), is regressed onto the specific factor (e.g. GHGs) to investigate the impacts of this factor on the TTL temperature variability. The results are consistent with the composite method presented in Figs. 3–7, which show positive temperature anomalies due to decreasing SSTs and increasing QBO amplitude, and negative temperature anomalies due to decreasing solar radiation and increasing GHGs in the TTL. The estimated contributions of Solar, SST, QBO and GHGs from the linear regression are, however, weaker than those from the composite method. The composite method is therefore a composite of "strong events", with remarkable decadal trend in these factors, and is thus useful to explain the recent decadal warming in the TTL.

4 Effects of the vertical resolution

To estimate not only anthropogenic and natural contributions to the recent TTL temperature variability but also the effects of the vertical resolution in the model, ~~Fig~~Figs. 8a and b ~~shows~~ show the temperature trends in the standard W_L66 run and the differences in temperature trends between the high-resolution (W_L103) and the standard (W_L66) runs, respectively. The W_L103 run (Fig. ~~7a~~8b) shows a statistically significant ~~warming~~ 0.6 $\mathrm{K\,decade^{-1}}$ warming on average over the past decade around the TTL, which maximizes at ~~1~~1.2 $\mathrm{K\,decade^{-1}}$. The trends in the W_L103 simulation in this paragraph were estimated by the linear regression method introduced in section 2.4, which is then applied to the three simulations of the W_L103 run. The resulting trends are almost the same as in Fig. 7b, but the significances are quite different because there are more sample sizes for the significance test. The standard W_L66 run (Fig. 8a) does not capture the warming. The only difference between the two experiments is the vertical resolution, meaning that a higher vertical resolution captures the warming in the TTL better than the standard vertical resolution, reaching up to ~~0.6 ± 0.2~~0.8 $\mathrm{K\,decade^{-1}}$ (Fig. 8b). ~~The uncertainty of the effect of vertical resolution is relatively small.~~ Wang et al. (2013) showed that the tropical upwelling in the lower stratosphere has weakened over the past decade in the W_L103 run, while there is no significant upwelling trend in the standard vertical resolution (W_L66) run. The decreasing tropical upwelling in the W_L103 run might be the reason for the extra warming in the TTL compared to the W_L66 run, since dynamical changes would lead to adiabatic warming. More detailed investigation will be given in the following section.

4.1 Changes in ~~wave-mean flow interaction~~the Brewer–Dobson circulation

To investigate dynamical differences between the two experiments with standard and higher vertical resolution in more detail, the Transformed Eulerian Mean (TEM) diagnostics (Andrews et al., 1987) ~~were~~ was applied to investigate differences in the wave propagation and ~~wave-mean flow interactions~~ Brewer–Dobson circulation (BDC) in the climatological mean as well as in the decadal trend. ~~The Eliassen–Palm flux (EP flux) vector shows the direction of wave propagation, and its divergence is a measure of wave-mean flow interactions. Figure 9 features the climatology of annual means of the EP flux vector (errors), the divergence of the EP flux vector (shading), and the zonal mean zonal wind (contours) from the WSUBSCRIPTNBL103 run (Fig. 9a), as well as the differences for each variable between the WSUBSCRIPTNBL103 and the WSUBSCRIPTNBL66 runs (Fig. 9b). Also shown are the trends for each variable in Fig. 9a from the WSUBSCRIPTNBL103 run (Fig. 9c) and the differences in the trends between the WSUBSCRIPTNBL103 and the WSUBSCRIPTNBL66 runs (Fig. 9d).~~

Figure ~~9a shows extratropical wave propagation upward from the lower to the upper troposphere and lower stratosphere in mid-latitudes, and an equatorward propagation of waves in the vicinity of the tropospheric subtropical jet, where some of the waves dissipate (divergence regions) (Fig. 9a). At the same time, equatorial waves propagate from the tropical lower troposphere to the tropopause region. In the finer vertical resolution~~

~~(WSUBSCRIPTNBL103) experiment, the upward EP flux from the mid-latitude troposphere is stronger than in the standard vertical resolution (WSUBSCRIPTNBL66) experiment in the middle to upper troposphere. The upward directed equatorial waves are also stronger because of the higher vertical resolution. Although upward propagation from both the mid-latitudes and tropics strengthens in the upper troposphere, waves are damped around the tropopause because the upper part of the subtropical jets is about 2.5weaker. Especially in the mid-latitude upper troposphere (12–16), there is a significant strengthening (weakening) of wave propagation from the extratropics to the tropics (from the tropics to the extratropics). At the same time, there is a stronger downward wave propagation in the tropical lower stratosphere (16–20). In summary, a finer vertical resolution leads to a stronger tropical upwelling from the lower to the upper troposphere (in the WSUBSCRIPTNBL103 as compared to the WSUBSCRIPTNBL66 experiment) in the climatological annual mean , but a weaker tropical upwelling in the lower stratosphere (see also discussion below).~~

~~The annual mean trend in the above discussed quantities is depicted in Fig. 9c for the high vertical resolution experiment. During the past decade significantly more waves propagated from the mid-latitudes into the tropics in the WSUBSCRIPTNBL103 experiment. This strengthening of wave propagation from higher latitudes to the tropics is most visible in the tropical upper troposphere between 11–17, and above the tropopause in the lower stratosphere above 18. There is also a trend towards slightly enhanced downward propagation around the TTL coinciding with a significant divergence of the EP flux in this region. A weaker upwelling in the mid to high latitude upper troposphere to lower stratosphere occurs in both hemispheres. A stronger upward propagation occurs in the tropical upper troposphere until about 12. Compared with the simulation in standard vertical resolution (run WSUBSCRIPTNBL66), the WSUBSCRIPTNBL103 run shows a stronger weakening of the upward wave propagation from the mid-to-high latitude upper troposphere to the lower stratosphere (Fig. 9d). In the tropical troposphere below 12, the strengthening of the upward propagation is much weaker in the run WSUBSCRIPTNBL103 compared with the WSUBSCRIPTNBl66 experiment in the Southern Hemisphere, but is stronger in the~~

~~Northern Hemisphere. How changes in the wave propagation affect the Brewer–Dobson circulation and hence temperature anomalies will be discussed in the following section.~~

4.2 ~~Changes in the Brewer–Dobson circulation~~

~~Figure 10 shows the annual mean~~ 9 shows the annual mean climatology of the ~~Brewer–Dobson circulation (BDC)~~ BDC (arrows for the meridional and vertical wind components), the zonal mean zonal wind (blue contour lines) and the temperature (filled colours) from the W_L103 run (Fig. ~~10a~~9a), as well as the differences between the W_L103 and the W_L66 runs (Fig. ~~10b~~9c). The BDC shows an upwelling in the tropics and a downwelling through mid to high latitudes in the annual mean. With finer vertical resolution (W_L103) the model produces a stronger upwelling in the tropics (and a consistent cooling) up to the ~~upper troposphere at around 16and in the tropical lower stratosphere. Around the tropical~~ tropopause region, with westerly wind anomalies above. This strengthened tropical upwelling can not continue further up because of the westerly wind anomalies blocking transport into the subtropics and finally diminishing. Above the tropical tropopause there is less upwelling and in particular more transport from the subtropics into the tropical TTL, leading to a stronger warming around 19 km in the W_L103 experiment. ~~The stronger and statistically significant warming in the mid-latitude troposphere around 13on both hemispheres corresponds to stronger transport from the tropics to the mid latitudes.~~ These changes in the BDC indicate a strengthening of its ~~transit branch (Lin and Fu, 2013)~~lower branch, and a weakening of the ~~upper branch~~transition branch (Lin and Fu, 2013). This is consistent with previous work by Bunzel and Schmidt (2013), which indicates a weaker upward mass flux ~~over~~ around 70 hPa in a model experiment with higher vertical resolution.

The annual mean trends in the W_L103 experiment indicate a further strengthening of the BDC ~~transit branch~~ lower branch over the past decade in this simulation (Fig. ~~10c~~9b) and a statistically significant weakening of the ~~upper branch~~ transition branch resulting in significant warming of 1 to 2 K decade^{-1} in the TTL. In ~~addition, there is a statistically significant weakening of the downwelling over the Southern Hemisphere high latitudes, which results~~

~~in a statistically significant cooling over the south polar region. In~~ particular the trends in the TTL are stronger in the W_L103 compared to the W_L66 experiment (Fig. ~~10d~~9d). This is consistent with previous work by Bunzel and Schmidt (2013), which shows also stronger changes in the BDC using a model with higher vertical resolution.

In summary, the finer vertical resolution can enhance the upward wave propagation from ~~both tropicsand extratropics~~the tropics. This enhanced wave propagation speeds up the ~~transit~~ lower branch of the BDC in the upper troposphere and slows down the ~~lower stratospheric~~ transition branch of the BDC. These changes in the BDC and corresponding wave-mean flow interactions (~~see above~~not shown) finally result in the statistically significant warming in the TTL.

Bunzel and Schmidt (2013) attributed the differences in the BDC to different vertical resolutions which tend to reduce the numerical diffusion through the tropopause and the secondary meridional circulation. Our results show that the strong warming and subsequent enhanced static stability (not shown) above the tropopause may also influence wave dissipation and propagation around the tropopause. ~~Due to~~ Oberländer et al. (2013) ~~, the strengthening of BDC from the past to the future is mainly caused by~~ point out that an increase of tropical SSTs ~~. Our results~~enhances the BDC. This is consistent with our results, which show a weakening of the ~~upper branch of the~~ BDC in the lower stratosphere following a ~~decline~~ decrease in tropical SSTs~~, and thereby support earlier work~~. At the same time, ~~a stronger warming in the TTL, due to a stronger weakening of the BDC in simulations with higher vertical resolution indicates that, with~~ this response of the stratosphere to the surface can be better represented by a model with finer vertical resolution~~, the model can also produce a better response of long-term variability in the TTL due to surface changes such as changes in tropical SSTs~~.

5 Summary and discussion

Based on a series of sensitivity simulations with NCAR's CESM-WACCM model, the contribution of different natural (solar, QBO, tropical SSTs) and anthropogenic (GHGs, ODS) factors to the observed warming of the TTL over the past decade from 2001 through 2011 has been studied. By comparing model experiments with and without the respective factors and combining a number of periods with similar trends in a composite, the contribution of each factor has been quantified in order to explain the causes of the observed recent decadal variability in GPS-RO data.

A decrease in tropical SSTs, an increase in stratospheric aerosol loading and an increase in the QBO amplitude contribute each about ~~0.3~~0.4, 0.4 and ~~0.3~~0.5 K decade^{-1} to this warming, respectively, resulting in a total ~~1.0~~1.3 K decade^{-1} warming, while the delay and smaller amplitude of the current solar maximum and the steady increase in GHGs and ODS concentrations contribute each about ~~0.2~~ 0.3 and 0.1 K decade^{-1} to a cooling, respectively, resulting in a total ~~0.3~~0.4 K decade^{-1} cooling. The vertical resolution of the model strongly influences the TTL response to the surface mainly via dynamical changes, i.e. an enhancement of the ~~(lower) transit~~ lower branch of the BDC and a decrease of the ~~upper~~ transition (upper) branch in response to the decreasing tropical SSTs. This leads to a ~~0.6~~0.8 K decade^{-1} extra warming in the TTL in the finer vertical resolution experiment as compared to the standard vertical resolution. ~~Summing~~ Adding all natural and anthropogenic factors, as well as the contribution from finer vertical resolution, we ~~arrive at~~ estimate a total modeled warming of ~~1.3~~1.7 K decade^{-1} around the TTL (Table 3), which is ~~in very good agreement with the observed 1.1~~higher than the observed 1.0 K decade^{-1} warming from GPS-RO data. ~~Although this is a very good agreement, care must be taken, since in reality non-linear interactions between the different factors might occur which we did not take into consideration in our first order linear approach. This uncertainty from the non-linear interactions should relatively small, which can be supported by our WSUBSCRIPTNBAerosol run.~~ The estimate from the model is higher because we estimated the total contribution from all factors and assumed that all factors are independent from each other and therefore the individual contributions can be linearly added. However, in reality non-linear interactions between the different factors occur which we did not take into account in our first order linear approach. This uncertainty from the non-linear interactions can be estimated by our W_Aerosol run. The W_Aerosol run, with almost all observed forcings considered in this study, can be seen as the most realistic simulation. The TTL warming in the W_Aerosol run is 1.0 K decade^{-1} on average and 1.6 K decade^{-1} in maximum (Fig. 7b), which ~~is~~ are very close to ~~both the observed and composite temperature trend.~~ the observed trend. This in turn indicates, that missing non-linear interactions, can overestimate the warming up to 0.7 K decade^{-1} compared to the observed 1.0 K decade^{-1} warming.

According to our experiments, one of the primary factors contributing to the recent warming in the TTL is the natural variability in tropical SSTs. ~~A change in the sign of tropical SST tendency, for example an SST increase from 1960 through 2000~~ (Randel et al., 2009) ~~and a decrease since 2001, leads to an opposite change in sign for TTL temperature tendency, which means a cooling from about 1960 to 2000 and a warming after 2001.~~ However, the mechanism ~~for this change in sign in the SSTs tendency, as well as the response~~ of the TTL response to SSTs awaits further investigation. ~~Our~~ One key issue is how much improvement we can expect from using a fully-coupled ocean-atmosphere model instead of atmosphere only model with prescribed SSTs. Our W
SUBSCRIPTNBL66 and W
SUBSCRIPTNBL103 simulations indicate that the ~~vertical resolution in the upper troposphere and lower stratosphere seems to be a key factor in simulating the~~ atmosphere-only model may not correctly reproduce the response of TTL variability to SST~~changes accurately~~, but can be improved with finer vertical resolution.

~~A similar change in sign can be seen in the trends of the QBO amplitude, which shows a decrease from about 1960 to 2000 but an increase since 2001. The long-term trend in QBO amplitude has been attributed by~~ Kawatani and Hamilton (2013) ~~to the variability in tropical upwelling near the tropical tropopause. However, the influences of this changing QBO amplitude on the temperature variability in the TTL is still unclear. Here, we made a simple estimation of this influence using a comparison between two simulations. There are still relatively large uncertainties in our estimation because of the nudged QBO in our model instead of a self-consistent QBO. The similar TTL temperature changes for long-term variability due to changes in tropical SSTs and QBO amplitude suggest that~~

~~they are~~ Another important factor in contributing to the recent warming in the TTL is the QBO amplitude. The QBO amplitude is closely related to ~~each other. A lagged correlation has been made for the time series of the QBO and the tropical SSTs from 1960 to 2000 (not shown). We find that a 0.6 correlation coefficient, with a lead time of about two years in tropical SSTs, which indicates that the SSTs may lead to the observed QBO long-term variability.~~ the tropical upwelling Kawatani and Hamilton (2013). A regression of temperature differences onto the differences in the vertical component of BDC between the Natural and NOQBO run, shows a very similar result than the regression of temperature onto the QBO time series (not shown). The QBO may influence the TTL temperature by modifying the BDC.

~~Another recent interesting signal of climate change can be found at the surface: the Earth's global average surface air temperature has stalled since around 2001 despite ongoing increases of atmospheric GHGs . The proposed causes for this strange behaviour include a decrease in solar variability, increasing water vapor concentrations or stratospheric aerosol loading, the strengthening of the Pacific trade winds and the enhanced heat transport from the surface to ocean depths over the past decade~~ (Balmaseda et al., 2013; England et al., 2014; Fyfe and Gillett, 2014; Kosaka and Xie, 2013)~~. The proposed causes for this strange behaviour include a decrease in solar variability, increasing water vapor concentrations or stratospheric aerosol loading, the strengthening of the Pacific trade winds and the enhanced heat transport from the surface to ocean depths over the past decade~~ (England et al., 2014; Kosaka and Xie, 2013; Fyfe et al., 2013)~~. A change in the sign of temperature tendency at around 2001, both at the surface and in the TTL, suggests that the surface and the TTL are closely related to each other. Understanding the mechanism for changes in the TTL can therefore help to understand the mechanisms for the global warming hiatus, but needs further detailed investigation.~~ Fig. S3 clearly shows decadal to multidecadal fluctuations in temperature in the TTL from both, the Modern Era Retrospective-analysis for Research and Applications (MERRA) reanalysis data, and our Natural and RCP85 runs, which provide a strong support to the internal variability dominated TTL warming over past decade.

The external forcings (solar, GHGs, ODS) contribute relatively little to the temperature variability in the TTL, except for the stratospheric aerosols. Internal variability, i.e. the QBO and tropical SSTs, seem to be mainly responsible for the recent TTL warming.

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Table 1. Overview of fully-coupled CESM-WACCM simulations (1955–2099).

Simulations	Natural Forcings	GHGs
Natural	All natural forcings, including transit solar variability, fully coupled ocean, prescribed volcanic aerosols and nudged QBO	Fixed GHGs to 1960s state
SolarMean	As Natural run, but with fixed solar radiation	Fixed
FixedSST	As Natural run, but with fixed SSTs	Fixed
NOQBO	As Natural run, but without QBO nudging	Fixed
RCP85	As Natural run	RCP8.5 scenario

Table 2. Overview of WACCM atmospheric stand-alone simulations (2001–2010).

Simulations	~~Ensemble Numbers~~ Number of Simulations	Vertical levels	Forcings	Stratospheric aerosols
W_L103	3	103	Observed solar variability and SSTs, nudged QBO, GHGs in RCP4.5 scenario	Volcanic aerosols from CCMVal-
W_L66	3	66	As W_L103	As W_L103
W_Aerosol	1	103	As W_L103	Stratospheric aerosols from CCM

Table 3. Summary of contributions from the varying factors to the observed TTL warming between 2001 and 2011, in the region 20° S–20° N latitude and 16–20 km.

Factors	Solar	SSTs	QBO	GHGs	Aerosols	Vertical Resolution	**Total**
Contribution (K decade^{-1})	-0.3	0.4	0.5	-0.1	0.4	0.8	**1.7**
Observation							**1.0**

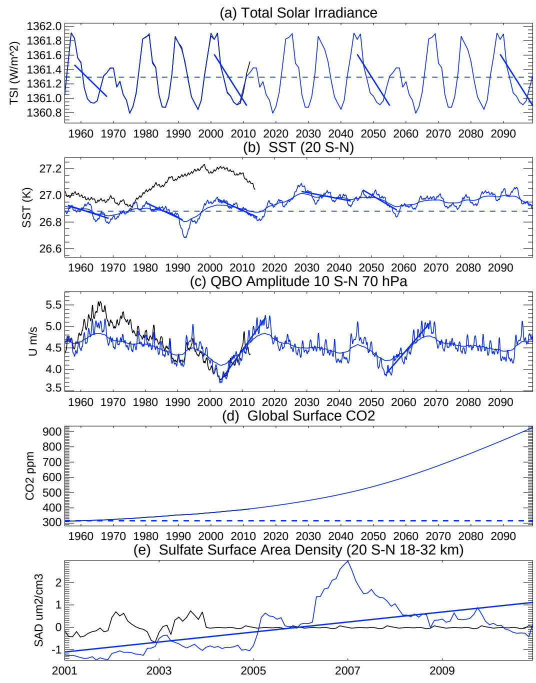
(a) Total Solar Irradiance
TSI (W/m^2)
(b) SST (20 S-N)
SST (K)
(c) QBO Amplitude 10 S-N 70 hPa
U m/s
(d) Global Surface CO2
CO2 ppm
(e) Sulfate Surface Area Density (20 S-N 18-32 km)
SAD um2/cm3

Figure 1. Time series of forcing data sets used for the simulations from 1955 through ~~2100.~~ 2099. **(a)** TSI from observations (black) until 2012, the *Natural* run (solid blue) and the *SolarMean* run (dashed blue). The last four solar cycles have been repeated into the future. **(b)** SSTs from observations HadISSTs (see details in text) (black), the *Natural* run (solid blue) and the *FixedSST* run (dashed blue). The SSTs in observations and the *Natural* run have been smoothed by a low-pass ($T > 30$ years) Butterworth Filter. The smooth blue line has been smoothed twice by the same low-pass Butterworth Filter. **(c)** Same as in **(b)**, but for the QBO amplitude calculated from zonal mean zonal winds at 70 hPa and between 10° S and 10° N. **(d)** Global surface CO_2 concentration from observations (black, overlapped with the blue line), the *RCP85* run (solid blue) and the *Natural* run (dashed blue). **(e)** AOD (532 nm, 18–32 km) from the CCMI (Solid blue) and the CCMVal2 (black) projects for the time 2001–2010. The blue solid straight lines in each subfigure are the linear fits of the respective forcing for the selected decade.

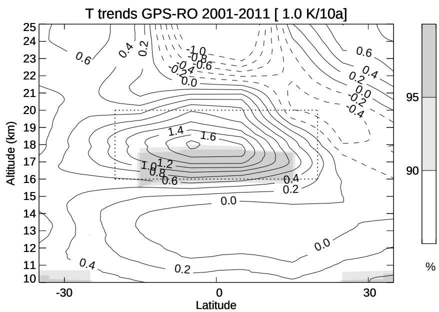

Figure 2. Latitude-height section of linear temperature trends over the past decade (2001–2011) from *GPS-RO* data over a height range from 10 to 25 km and 35° S to 35° N latitude; contour interval: ~~0.3~~0.2 K decade^{-1}. Grey shading represents the statistical significance for the trends. See text for details on the linear trend and the statistical significance calculation.

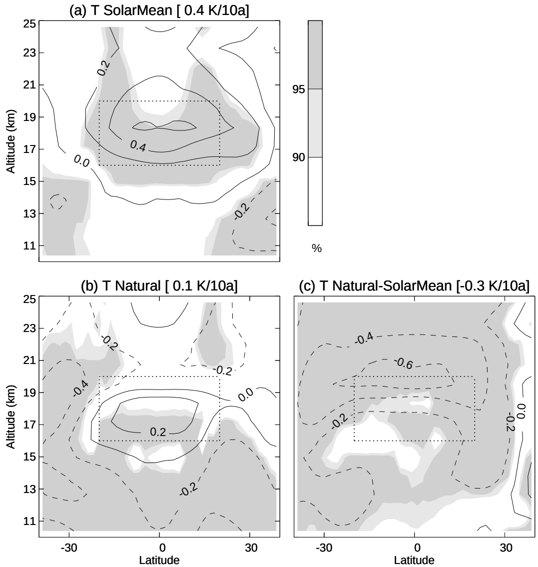

Figure 3. (a, b) Latitude-height sections of composite temperature trends over selected time periods (1958–1968, 2001–2011, 2045–2055, and 2089–2099, see Fig. 1) from the *Natural* and *SolarMean* runs, respectively; contour interval: 0.2 K decade^{-1}. **(c)** The differences between **(a)** and **(b)**. Grey shading represents statistically significant trends~~(differences)~~. See text for details on the calculation of the composite trend ~~and the trend differences~~, and the testing of the statistical significance~~for both of the composite~~. The decadal temperature trend ~~and~~ in the ~~trend differences~~title is the mean value from the dashed box.

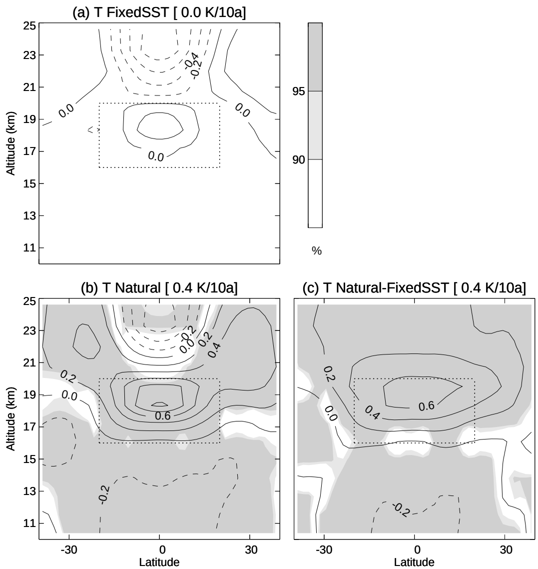

Figure 4. (a, b) Latitude-height sections of composite temperature trends over selected time periods (1956–1968, 1980–1991, 2001–2014, 2028–2043, 2047–2057, see Fig. 1) from the *Natural* and *FixedSST* runs, respectively; contour interval: 0.2 K decade^{-1}. **(c)** The differences between **(a)** and **(b)**. Grey shading as in Fig. 3.

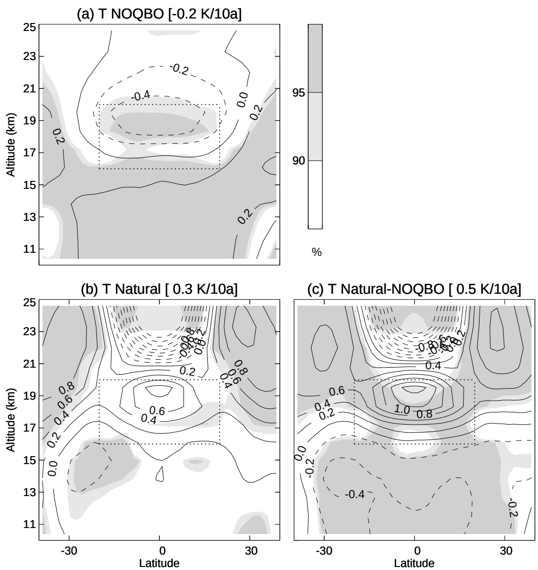

Figure 5. Same as Fig. 3, but for the impact of the QBO amplitude on temperature trends **(c)** by comparing the *Natural* and the *NOQBO* experiments **(a, b)** for the periods 2003–2017 and 2054–2068 (see Fig. 1); contour interval: 0.2 K decade^{-1}.

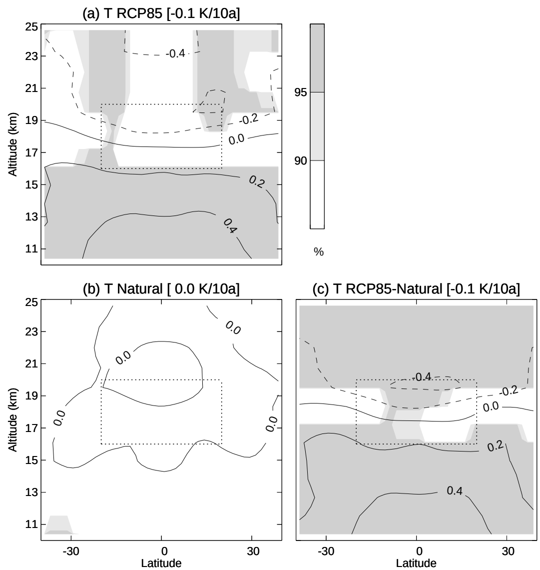

Figure 6. Same as Fig. 3, but for the impact of anthropogenic forcings (GHGs and ODS) on temperature trends by comparing the *Natural* and *RCP85* experiments **(a, b)** for the period ~~2001–2011~~2001–2050; contour interval: 0.2 K decade^{-1}.

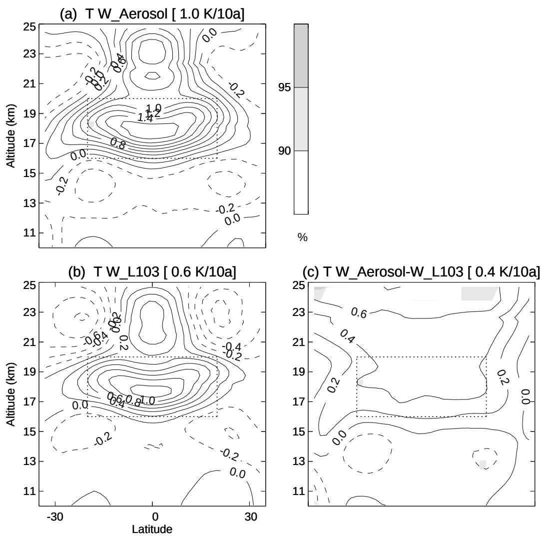

Figure 7. Same as Fig. 3, but for the impact of stratospheric aerosols on temperature trends by comparing the *W_L103* and the *W_Aerosol* experiment **(a, b)** for the period 2001–2010; contour interval: 0.2 K decade^{-1}. The temperature trends in the *W SUBSCRIPTNBL103* run were calculated by ensemble mean.

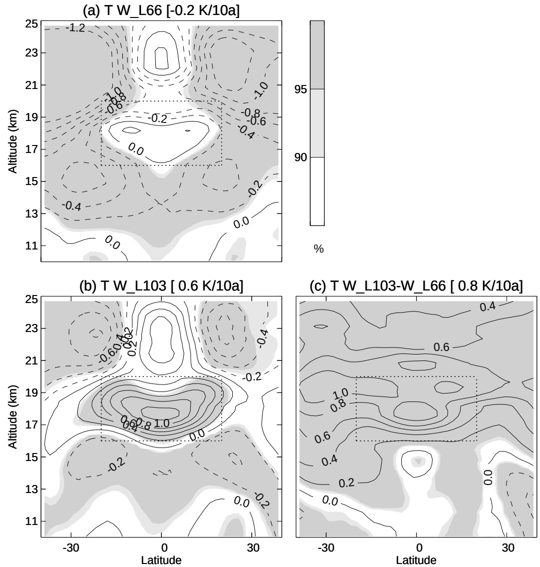

Figure 8. ~~**(a)** Latitude-height sections of temperature trends between 2001 and 2010 from~~ Same as Fig. 3, but for the ~~ensemble of the *WSUBSCRIPTNBL66* experiment to investigate the~~ impact of the differences in vertical resolution on temperature trends **(c)** by comparing the *W SUBSCRIPTNBL103* and *W SUBSCRIPTNBL66* experiments **(a, b)** for the period 2001–2010; contour interval: 0.2 K decade^{-1} and grey shading as in Fig. 3. ~~**(b)**~~ The ~~differences between (Figs. 7a~~temperature trends in the *W SUBSCRIPTNBL103* and ~~8a)~~ *W SUBSCRIPTNBL66* runs are calculated by multiple linear regression for the three ensembles.

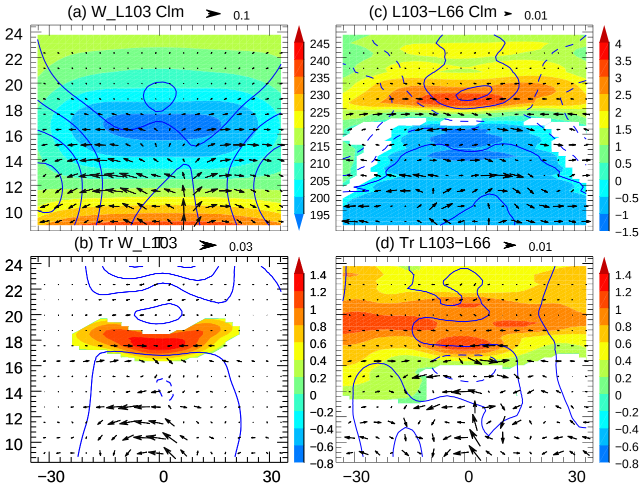

Figure 9. **(a)** Annual mean climatological zonal mean zonal wind (contours, contour interval ~~5~~10 m s^{-1}, dashed lines indicate easterly winds~~, the thick line is the zero wind line~~), ~~EP flux~~ BDC vector (arrows, scaled with the square root of pressure) and ~~its divergence~~ temperature (~~shading, positive values indicate divergence~~colour shadings) for the *W_L103* experiment from 8 to 25 km and ~~80~~35° S through ~~80~~35° N. **(c)** Differences of the zonal mean zonal wind (contour interval ~~0.5~~1.0 m s^{-1}), ~~EP flux~~ BDC vector and ~~its divergence~~ temperature (colour shadings indicate 95 % statistical significances) between the *W_L103* and the *W_L66* experiments. (**b** and **d**) Same as **(a)** and **(c)**, but for the linear trends from 2001 to 2010. The shadings in **(b)** ~~indicate 95statistical significance. The shadings in~~ and **(d)** indicate 95 % statistical significance~~in both **(b)** and **(c)**~~. The contour intervals are ~~1~~2 m s^{-1} and ~~0.5~~1 m s^{-1} in **(c)** and **(d)**, respectively.

~~Same as Fig. 9, but for the vertical and meridional components of the residual circulation (arrows) and the zonal mean temperature (shading).~~