Shift in cold-point tropopause trends derived from radiosonde, satellite, and reanalysis data

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Abstract.

The tropical tropopause layer (TTL) is the transition region between the well-mixed convective troposphere and the radiatively controlled stratosphere and plays a crucial role for air mass transport between these layers. In this paper, we present updated trends of TTL and lower stratospheric temperature from radiosonde and Global Navigation Satellite System – Radio Occultation (GNSS-RO) data and evaluate temperature trends in the reanalysis data sets ERA5, JRA-3Q, and MERRA-2. Given its importance in determining the concentration of water vapor entering the stratosphere, we focused in particular on temperature trends at the cold point tropopause, which we determined from radiosonde observations by removing time-varying bias effects from trends based on unadjusted data.

From 1980 to 2023, cold point tropopause cooling is shown in radiosondes, overestimated by JRA-3Q and underestimated by MERRA-2 and ERA5. Splitting into two periods reveals a shift in TTL temperature trends: cooling (1980-2001) to warming (2002-2023) across all datasets, highlighting post-2002 changes in tropical tropopause dynamics. Tropical upwelling estimates from the three reanalyses show opposite trends for 1980-2001 compared to 2002-2023 consistent with the cold point and lower stratosphere temperature trends. While the vertical residual circulation increased before 2000 consistent with cold point cooling, the circulation trends turned to zero (MERRA-2, JRA-3Q) or became positive (ERA5) after 2000 consistent with cold point warming. Between 2002-2023, GNSS-RO and reanalysis show significant warming at the cold point and lower stratosphere, aligning with observed patterns and seasonality. Warming trends anticorrelate with tropospheric cooling, strongest where upper troposphere cooling appears.

1 Introduction

The tropopause, acting as a boundary between the troposphere and stratosphere, plays a crucial role for air mass transport between these layers. Its variations can significantly impact atmospheric dynamics, circulation, and the distribution of greenhouse gases in the upper troposphere and lower stratosphere (UTLS, Holton et al., 1995; Stohl et al., 2003). Therefore,

quantifying the tropopause's climatological features and trends is vital for understanding and simulating these broader atmospheric processes. In the tropics, the cold-point tropopause (CPT) is identified as the point of minimum temperature in the vertical temperature profile (Highwood and Hoskins, 1998). The CPT is particularly important for stratospheric composition and regulating water vapor entry into the stratosphere, making it essential to study both its properties and variations.

Numerous studies have examined the variability and trends of the tropopause layer using various data sources, including radiosonde measurements, satellite observations, reanalysis data, and climate models (Randel et al., 2006; Seidel and Randel, 2006; Rosenlof and Reid, 2008; Wang et al., 2012; Randel et al., 2017; Zolghadrshojaee et al., 2024). These studies have contributed to our understanding of tropopause behavior and its implications for atmospheric processes. Focusing on the time period before mid-2000s, studies have documented a long-term cooling of the cold point temperature. Rosenlof and Reid (2008) reported a cooling trend at the cold point and at 100 hPa between 1980 and 2003, with temperature decreases ranging from −0.5 to −1 K per decade, based on radiosonde observations from 52 stations. This finding aligns with Zhou et al. (2001), who observed a cooling trend of about −0.57 ± 0.06 K per decade of the CPT from 1973 to 1998 using operational sounding data. Focusing on the tropical region (15°S-15°N), Seidel et al. (2001) analyzed data from 83 radiosonde stations for the period 1978-1997. They found that the tropical tropopause height increased by 20 meters per decade, while its temperature decreased by −0.5 K per decade. The reliability of these observations is supported by studies comparing different measurement techniques. Anthes et al. (2008) and Ho et al. (2017) demonstrated strong consistency between temperature profiles obtained from the Global Navigation Satellite System – Radio Occultation (GNSS-RO) and those from radiosondes, lending credibility to the observed trends. However, Wang et al. (2012) demonstrated that adjusted radiosonde temperatures give smaller cooling trends for 1979-2005 with larger uncertainties than the unadjusted data sets. In addition to the overall cooling trend before the early 2000s, tropopause temperatures appear to have dropped abruptly by ∼1.5° to unusually and persistently low values in the years after 2000, consistent with a similar drop in stratospheric water vapour (Randel et al., 2006; Rosenlof and Reid, 2008). Recent studies have revealed a shift in tropical tropopause temperature trends. In contrast to earlier cooling trends, Wang et al. (2015) observed a warming trend in the tropical tropopause layer (TTL) from 2001 to 2011. Their analysis of GNSS-RO data showed a temperature increase of 0.9 K per decade. Similarly, Randel and Park (2019) identified warming of the CPT from 2010 to 2017, consistent with increased water vapor in the tropopause region as detected by NASA's Aura satellite Microwave Limb Sounder (MLS). Extending the analysis of GNSS-RO data to 2002-2022, Zolghadrshojaee et al. (2024) confirmed the cold point warming with the strongest warming during the boreal winter and spring reducing the amplitude of the cold-point temperature and water vapour seasonal cycles. The above findings suggest a significant change in tropopause behavior: cooling trends from the 1970s and 1980s until the early 21st century, followed by a transition to warming in recent years. The representation of changes of tropical tropopause temperature in meteorological reanalyses is of interest for studies of stratospheric transport, composition, and long-term changes. Reanalysis data sets are widely used in such scientific studies, often as "stand-ins" for observations, when the available measurements lack the spatial or temporal coverage

needed. In particular, reanalysis products in the TTL region are important for transport and composition studies with offline
chemical transport models or Lagrangian particle dispersion models (e.g., Chipperfield, 1999; Krüger et al., 2009; Schoeberl
et al., 2012; Tao et al., 2019). The representation of the cold point in the reanalysis data set used to drive these models
determines how realistically such models simulate dehydration and stratospheric entrainment processes.

Reanalysis applications can be complicated by spurious changes and discontinuities in the reanalysis fields due to changes in
the quality and quantity of the observations used as input data or the joining of different execution streams (e.g., Fujiwara et
al., 2017). A comparison of reanalysis products available at the end of the 1990s with other climatological data sets showed
notable differences in temperatures near the tropical tropopause (Randel et al., 2004). Reanalysis cold point temperature and
height data used by Gettelman et al. (2010) for the evaluation of model results showed a considerable spread of the identified
cooling trends. Advances of the reanalysis and observational systems over the last decades have led to a clear improvement in
the TTL reanalysis products over time with smaller biases and better agreement of the variability from 2006 onward, when
densely sampled radio occultation data started being assimilated by the reanalyses (Tegtmeier et al., 2020). The authors
reported that most reanalyses suggest small but significant cold point cooling trends of −0.3 to −0.6 K per decade for 1979-
2005 that are statistically consistent with trends based on the adjusted radiosonde data sets.

We aimed to update existing studies of the variability and trends of tropopause temperatures from various observational and
reanalysis datasets covering an extended period from 1980 to 2023. Our analysis will focus on providing adjusted cold point
temperature trends based on radiosonde data for the full time period. Detailed analysis of the seasonal cycle and spatial patterns
of the temperature trends are given for the more recent 2002-2023 time period based on GNSS-RO data. Reanalysis
temperature trends will be evaluated for all the above diagnostics. Sections 2 and 3 describe the datasets utilized and the applied
methodology. In Section 4, we presented recent temperature trends for 2002-2023, while Section 5 focuses on trends over
1980-2001.

2 Data

2.1 Unadjusted radiosonde data

We used a monthly mean, gridded temperature dataset at the cold point as well as at the 100 and 70 hPa pressure levels based
on the unadjusted, quality-controlled Integrated Global Radiosonde Archive version 2.2 (IGRA 2.2; Durre et al., 2006) for the
period from 1980 to 2023. IGRA data stems from radiosonde and pilot balloon observations from over 2,800 globally
distributed stations with the first data from soundings in 1905.

For calculating the cold point based on unadjusted IGRA data, we followed the criteria outlined by Wang et al. (2012). For
each sounding at one of the 110 tropical stations, we selected the minimum temperature across six pressure levels ranging from
300 hPa to 70 hPa (specifically, 300, 250, 200, 150, 100, and 70 hPa), excluding soundings that lacked data at any level
between. In a second step, monthly mean cold-point temperatures at each station were calculated for T00 (00:00 UTC) and
T12 (12:00 UTC) requiring at least 15 daily recordings within a month for each time respectively.

Trends of cold-point as well as 70 and 100 hPa temperatures were derived from monthly mean data based on the multiple linear regression analysis (Section 3.1), provided that at least 70% of the data was available. Outliers were identified and removed using the Interquartile Range (IQR) method. Data points falling outside the 25th and 75th percentiles of the temperature distribution for each station (less than 1% on average) were considered outliers and excluded from the dataset prior to trend calculation. Trends were computed for each station individually and then averaged over each grid cell of a 30° longitude × 16° latitude grid within 30°S to 30°N to obtain mean trends for T00 and T12 separately. In the final step, tropical average (30°S-30°N) temperature trends were obtained by averaging the trends over all grid cells and both times. The dataset consists of observations from 144 stations, some of which have incomplete records due to gaps in their time series. Additionally, the spatial distribution of stations is uneven, with substantial gaps over the oceans.

2.2 Adjusted radiosonde data

The accuracy of temperature trends derived from radiosonde data can be impacted by inconsistencies or biases resulting from changes in instrumentation or measurement methodologies (Seidel and Randel, 2006). To address potential biases in trends estimated from unadjusted radiosonde data, we utilized two adjusted radiosonde temperature datasets, the Radiosonde Observation Correction Using Reanalysis (RAOBCORE; Haimberger, 2007), and Radiosonde Innovation Composite Homogenization (RICH; Haimberger et al., 2008). RAOBCORE adjustments are based on reanalysis used to locate and adjust temporal discontinuities, while RICH determines adjustments for the same change points by using neighboring stations. These adjusted gridded data sets (10° longitude × 10° latitude) are available only at widely spaced fixed pressure levels (100 and 70 hPa in the upper TTL), and not at the cold point. The adjusted data sets come at monthly resolutions spanning the period from 1980 to 2023 and do not provide information on individual soundings.

Deseasonalized tropical mean temperature anomalies for the CPT, as well as at the 70 hPa and 100 hPa levels based on IGRA, RAOBCORE, and RICH datasets from 1980 to 2023 are shown in Fig. 1. Anomalies are calculated relative to the mean annual cycle for 1980-2023 for each data set. The 70 hPa level (Fig. 1a) shows slightly larger temperature variations compared to the other two levels, with anomalies ranging approximately from -2K to +3K. All three datasets show very close agreement of their interannual anomalies, which are dominated by the stratospheric QBO signal. The situation is very similar at the 100 hPa level (Fig. 1c) with excellent agreement between RAOBCORE and RICH and relatively good agreement of IGRA with the other two. At the cold point, only IGRA data is available and displays interannual anomalies and long-term changes similar to the signals at 70 and 100 hPa. Positive temperature anomalies following the eruptions of El Chichón in 1982 and Mount Pinatubo in 1991 can be detected at the 70 hPa level but are less evident at the cold point and the 100 hPa level (Fujiwara et al., 2015).

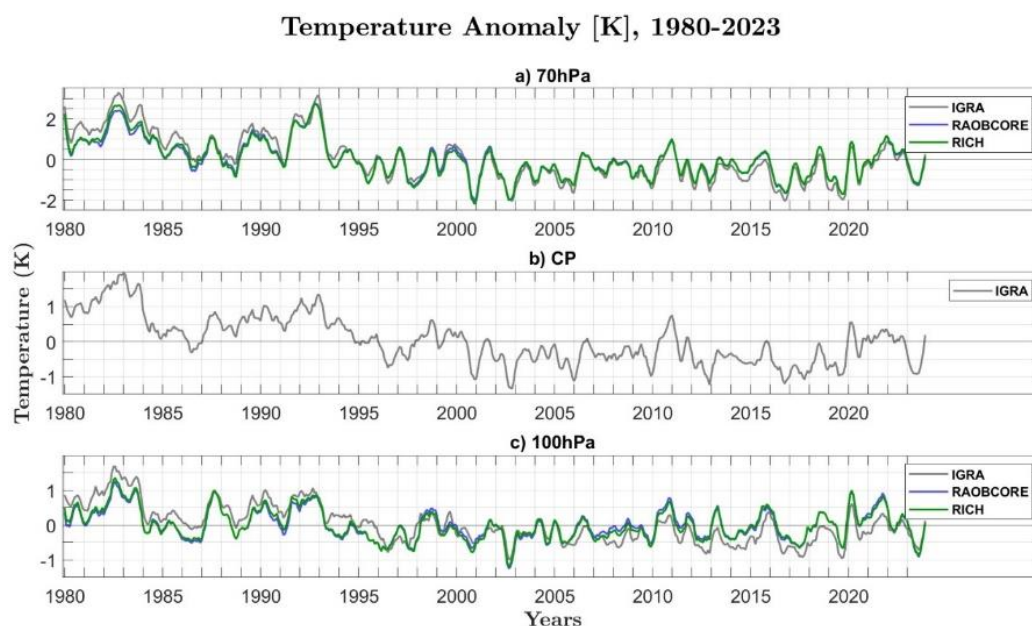



Figure 1. Tropical mean (30°S-30°N) deseasonalized temperature anomalies from 1980 to 2023 for (a) 70 hPa, (b) the cold-point tropopause (CPT), and (c) 100 hPa levels, using IGRA (gray), RAOBCORE (blue), and RICH (green) datasets. The data are
130 **smoothed over 5 months for visibility.**

2.3 Satellite data

We utilized GNSS-RO data, which offers global coverage, remains unaffected by clouds and precipitation, and has high vertical resolution and accuracy (Kursinski et al., 1997). In this study, we combined various satellite missions, including
135 CHAMP (Wickert et al., 2001), COSMIC (Anthes et al., 2008), GRACE (Beyerle et al., 2005), Metop-A (Von Engeln et al., 2011), successive Metop-B, METOP-C, SAC-C (Hajj et al., 2004), TerraSAR-X (Beyerle et al., 2011), and KOMPSAT-5 (Bowler, 2018). Specifically, we used monthly mean CPT temperature, along with temperature values at fixed pressure levels (30, 50, 60, 70, 100, 150, 200, 300, and 400 hPa). These data are sourced from all GNSS-RO missions and are spatially gridded on a 30° longitude × 10° latitude grid within the latitudinal range of 30°N to 30°S, covering the period 2002-2023. We
140 calculated temperature trends at the cold point and at fixed pressure levels for the gridded data and for tropical mean averaged data following the methodology given in Section 3.1.

145 ## 2.4 Reanalysis data

We used temperature data from three reanalysis datasets, the Modern-Era Retrospective Analysis for Research and Applications, v2 (MERRA-2; Gelaro et al., 2017), the ECMWF Reanalysis v5 (ERA5; Hersbach et al., 2020), and the Japanese

Reanalysis for Three-Quarters of a Century (JRA-3Q; Kosaka et al., 2024) covering the period from 1980 to 2023. Our analysis is based on gridded monthly mean tropical temperature fields extracted at the cold point and at various pressure levels encompassing the TTL and extending into the upper troposphere and lower stratosphere.

Lower stratospheric temperature from ERA5 has a cold bias for the years from 2000 to 2006 as a result of the specified background error covariances for the data assimilation. These covariances were chosen inappropriately for the period starting with the introduction of AMSU data in 2000 and ending with the availability of large amounts of GNSS-RO data in 2006, which successfully constrained the cold bias of the assimilating model. The bias was corrected in a new product (ERA5.1) produced for 2000-2006, which provides improved global-mean temperatures in the stratosphere and uppermost troposphere (Simmons et al., 2020). Here we used a combined product of ERA5.1 for 2000-2006 and ERA5 for all other years, which we refer to as ERA5 for simplicity.

Conventional observations—including surface measurements, balloon readings, aircraft data, and satellite observations—are assimilated to constrain the temperature fields in the reanalysis data. In the following, we listed the most important assimilated observations with further details given in Tegtmeier et al. (2020). All of the above reanalysis systems assimilate microwave and infrared radiance from the satellite sounders of the TOVS suite (1979–2006) and the ATOVS suite (1998–present). The introduction of ATOVS measurements around 1998, with a higher number of channels compared to TOVS, considerably improved the vertical resolution of the assimilated data. In addition, MERRA-2, ERA5, and JRA-3Q assimilate radiance estimates from the hyperspectral infrared sounders AIRS (2002–present), IASI (2008–present), and/or CrIS (2012–present). To ensure that effects of radiance biases associated with instrument changes, calibration offsets, orbital drifts, or long-term CO_2 changes are kept small, variational bias correction schemes are used during the data assimilation procedure.

Temperature observations from radiosondes are assimilated by all reanalyses, with systematic errors due to solar heating of the temperature sensor typically being corrected either on-site or at the reanalysis center before assimilation (Fujiwara et al., 2017). By assimilating RAOBCORE (JRA-3Q, MERRA-2) and RICH (ERA5) data discontinuities in temperature time series from unadjusted radiosondes are avoided.

The reanalysis models resolve the TTL with different vertical resolutions, with the number of model levels between 200 and 70 hPa varying from 14 (JRA-3Q) and 7 (MERRA-2) to 21 (ERA5). These model-level resolutions correspond to vertical resolutions between ~1 and ~0.2km. The horizontal resolutions of the reanalysis products are approximately 0.25°× 0.25° (ERA5), 0.667° × 0.5° (MERRA-2), and 0.375° × 0.375° (JRA-3Q). The cold point tropopause temperature was computed by taking the minimum temperature for each reanalysis using model-level data between 500 and 10 hPa at each grid point and at the finest temporal resolution available for model level output (i.e., hourly for ERA5, 3-hourly for MERRA-2, and 6-hourly for JRA-3Q). The data used for the CPT calculation are at the native output horizontal resolution of each reanalysis except for ERA-5, which was regridded to 1° x 1° as part of the data download due to bandwidth and file size constraints. The reanalysis model-level pressure fields were interpolated to fixed pressure levels (30, 50, 60, 70, 100, 150, 200, 300, and 400 hPa) at each

180 grid point. For JRA3Q, the nominal model-level monthly mean data were used, while for MERRA-2 and ERA5, the monthly mean pressure-level data were directly employed in the analysis.

Zonal and tropical averages of the cold point and pressure level temperature fields were calculated by averaging over all grid points, representing the final step of data processing.

We also employed variables from the zonal-mean data set of global atmospheric reanalyses on pressure levels provided by
185 Martineau et al. (2018). In particular, we used the vertical residual velocity, $\overline{\omega}^*$, from the Transformed-Eulerian-mean diagnostics based on the primitive momentum equation. Vertical velocities in reanalysis products are known to be highly dependent on the specific implementation of data assimilation step and less constrained by observations than the meridional velocities. A comparison of the climatological vertical residual velocities from various reanalyses to the corresponding estimates based on the meridional residual velocity through the continuity equation revealed reasonable agreement between
190 the two vertical velocity fields in the troposphere and lower stratosphere up to 10 hPa (Fujiwara et al., 2024).

3. Methods

3.1 Multiple linear regression and trend error bars

We used multiple linear regression to distinguish between long-term trends and short-term variations in TTL temperatures,
195 accounting for various atmospheric phenomena. This analysis utilized deseasonalized temperature anomalies at both the CPT and standard pressure levels, using gridded data and tropical averages from the given datasets listed above. By applying the regression described in Eq. (1), we separated the gradual temperature changes over time from those influenced by factors such as QBO, ENSO, and stratospheric aerosol optical depth (SAOD).

$$T(t) = A + B \cdot t + C_1 \cdot QBO_1(t) + C_2 \cdot QBO_2(t) + D \cdot ENSO(t) + E \cdot SAOD(t) + \epsilon(t), \qquad (1)$$

200 where $T(t)$ are the temperature anomalies and $\epsilon(t)$ are the residuals. The output of the regression provides the linear trend in the form of the regression coefficient B. The $QBO_1(t)$ and $QBO_2(t)$ terms are orthogonal time series representing QBO variations constructed as the first two empirical orthogonal functions of the Freie Universität Berlin (FUB) radiosonde stratospheric winds (Naujokat, 1986; Wallace et al., 1993). $ENSO(t)$ is the multivariate ENSO index (https://www.esrl.noaa.gov/psd/enso/mei/), and $SAOD(t)$ is the detrended stratospheric aerosol optical depth from the Global
205 Space-based Stratospheric Aerosol Climatology (Kovilakam et al., 2020).

We calculated the uncertainty in each long-term trend as the standard error of the slope with the effective sample size adjusted to account for the corresponding lag-1 autocorrelation coefficient as described in Santer et al. (2000). Significance was tested based on a two-tailed test with a 95 % confidence interval.

210 ### 3.2 Nearby Level approach

Since the adjusted radiosonde datasets lack cold-point temperature, we employed the methodology described by Wang et al. (2012) to generate CPT temperature trends for these datasets. Specifically, we applied the 'Nearby Level' approach, which compares temperature trends at the CPT and nearby fixed pressure levels in unadjusted data with trends at those fixed levels in the two adjusted datasets. This approach makes use of the fact that inhomogeneities at the cold point usually occur at the

215 same time as those at the surrounding levels as discussed by Wang et al. (2012). Therefore, the trend of the time series of differences can be assumed to be mostly independent of the inhomogeneities. In addition, the 'Nearby Level' approach assumes that the CPT trend can be inferred from the adjusted trend at a specific pressure level.

In the first step, the tropical mean temperature trends were calculated for unadjusted data at the cold point, 100 and 70 hPa, and for adjusted data at 100 and 70 hPa. In the second step, we applied the Nearby Level approach by combining the fixed

220 level trend (using adjusted data) with the trend in the differences between cold point and fixed level (using unadjusted data), as described in Eq. (2).

$$Trend_{CP,adj} = Trend_{fixed,adj} + Trend(T_{CP,unadj} - T_{fixed,unadj}), \qquad (2)$$

where $Trend_{fixed,adj}$ is the trend of adjusted datasets (RAOBCORE or RICH) at the fixed pressure level (70 or 100hPa), while $T_{CP,unadj}$ and $T_{fixed,unadj}$ are the unadjusted temperatures (IGRA) at the cold point and 70 or 100hPa, respectively.

225 Figure 2 shows the tropical average temperature trends from 1980 to 2023 before and after applying the Nearby Level approach. The adjusted radiosonde datasets suggest statistically significant cooling at the lower stratospheric level of 70 hPa of around -0.39 ± 0.15 K per decade (Fig. 2a). At the 100 hPa level, only very weak and non-significant cooling trends are derived from the adjusted radiosondes. The unadjusted IGRA data suggests stronger cooling at both levels, most likely due to spurious cooling introduced by effects such as the solar heating of the temperature sensors. These stronger cooling signals are

230 also apparent in the anomaly time series (Fig. 1), where IGRA shows more positive anomalies at the beginning of the time period and more negative anomalies at the end of the time period when compared to the other two data sets.

Given that the cold-point trend based on unadjusted IGRA data (Fig. 2a) is not reliable, the Nearby Level approach was applied to derive adjusted cold-point temperature trends (Fig. 2b) ranging from -0.18 ± 0.10 K per decade (based on 100 hPa RAOBCORE) to -0.21 ± 0.16 K per decade (based on 70 hPa RICH data). The good agreement of the four trend estimates

235 highlights the consistency of the CPT temperature trends derived from adjusted data sets, which are only half of the cooling trend suggested by unadjusted data.

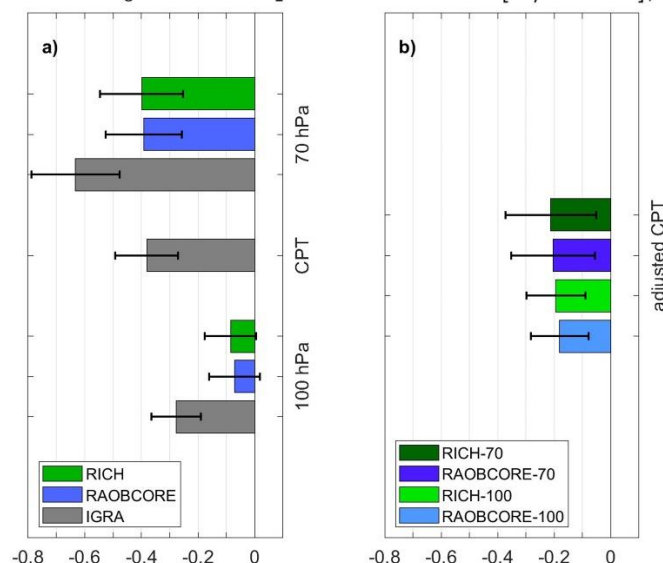



Figure 2. Tropical average (30°S-30°N) temperature trends during 1980-2023 for (a) the cold-point tropopause (CPT) and nearby pressure levels for unadjusted (IGRA) and adjusted (RAOBCORE, RICH) data and (b) for the CPT trends based on adjusted data and the Nearby Level approach. Error bars show the ± 2 sigma confidence intervals of the trend taking into account the adjusted effective sample size.

4. Temperature trends for 2002 to 2023

Cold-point temperatures experienced a drop-like decrease around the year 2000 and remained anomalously cold over the following years (Randel et al., 2006). This feature is also known from stratospheric water vapor changes, which are highly correlated with cold-point tropopause temperatures. GNSS-RO temperature data became available from 2002 onward allowing for more detailed studies of TTL temperature features and trends. In order to avoid the impact of the temperature drop in 2000 on trend estimates and to make use of the high-resolution, dense GNSS-RO data, we first focused our analysis on the 2002-2023 time period and compared the various observational and reanalysis data sets. Figure 3 shows the tropical mean (30°S-30°N) CPT temperature anomalies from observations (GNSS-RO) and reanalysis (MERRA-2, JRA-3Q, and ERA5). The time series shows large amounts of inter-annual variability, which is well captured across the reanalysis datasets when compared to the satellite data. Agreement among the reanalyses and between the reanalyses and the observations generally improved in 2006 with the beginning of the assimilation of COSMIC GNSS-RO data. Despite the large inter-annual fluctuations, a general warming trend is evident. The consistency across all four datasets strengthens the conclusion that tropical CPT temperatures have been increasing over the past two decades, highlighting the importance of ongoing atmospheric monitoring and analysis.

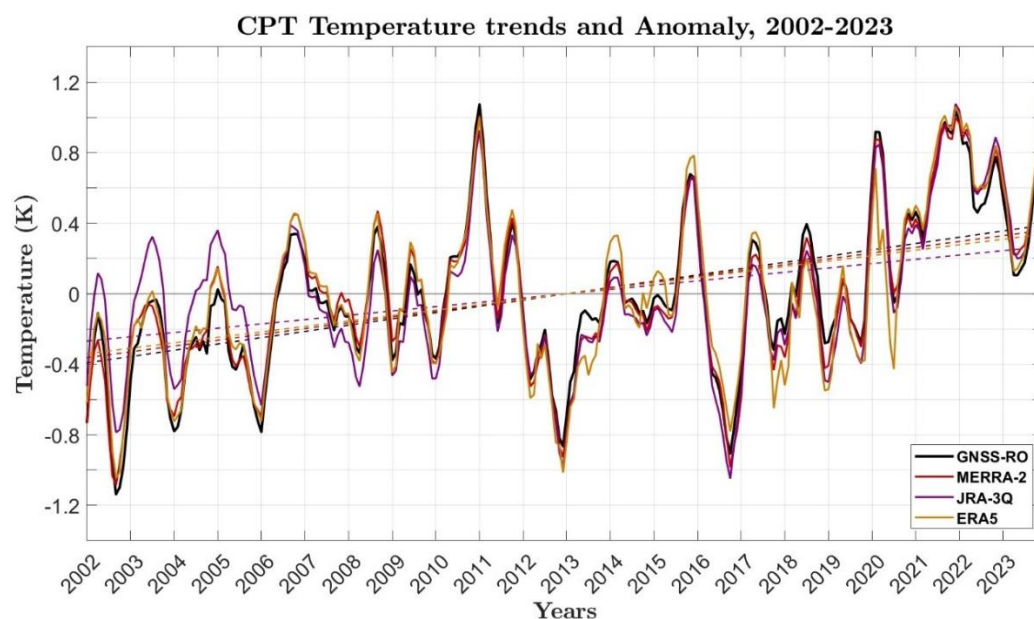



Figure 3. Tropical mean (30°S-30°N) deseasonalized cold-point tropopause (CPT) temperature anomalies and the linear trend based on the GNSS-RO, MERRA-2, JRA-3Q, and ERA5 data. The data are smoothed over 5 months for visibility.

The cold-point temperature trends for 2002-2023 for the three reanalysis and GNSS-RO data together with the radiosonde trends are shown in Fig. 4. The reanalysis trends agree very well with the GNSS-RO trend (Fig. 4c), with all four data sets showing significant warming which is strongest for MERRA-2 (0.33 ± 0.17 K per decade) and weakest for JRA-3Q (0.25 ± 0.19 K per decade). These warming trends seem to be driven by the appearance of particularly low-temperature anomalies at the beginning of the time period (2002 and 2006) and particularly high-temperature anomalies at the end of the time period (2020-2023) as illustrated in Fig. 3.

The adjusted radiosonde trends (Fig. 4b) also suggest warming, however with non-significant trend estimates. All four estimates are very similar ranging from 0.14 ± 0.27 K per decade (based on 100 hPa RAOBCORE data) to 0.23 ± 0.39 K per decade (based on 70 hPa RICH data). These trends are about half as strong as the satellite-based warming trend (GNSS-RO) of 0.36 ± 0.16 K per decade consistent with results from Steiner et al., (2020) for the time period 2002-2018. Sampling the reanalysis according to the radiosonde station network does not change the reanalysis trend estimates (not shown here) implying that the limited sampling of the radiosonde network is not the reason for the smaller trend estimates.

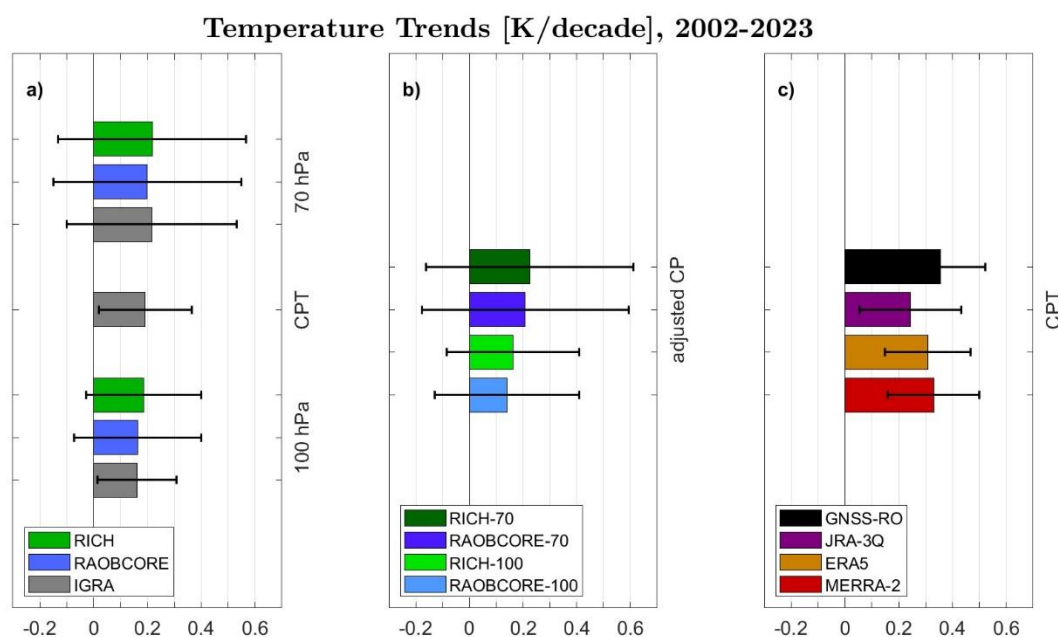



Figure 4. Tropical average (30°S-30°N) temperature trends during 2002-2023 for (a) the cold-point tropopause (CPT) and nearby pressure levels for unadjusted (IGRA) and adjusted (RAOBCORE, RICH) data; (b) the CPT trends based on adjusted data and the Nearby Level approach and (c) the CPT based on GNSS-RO and reanalysis data. Error bars show the ± 2 sigma confidence intervals of the trend taking into account the adjusted effective sample size.

We analyzed the spatial patterns of the cold point temperature trends using regression analysis applied to monthly temperature time series at each grid point for satellite observations and reanalysis (Fig. 5). All datasets show a general warming trend in CPT temperatures across the tropics. Pronounced warming can be seen across all data sets over the tropical North Atlantic Ocean, tropical South Pacific Ocean, and South Indian Ocean. Temperature trends are insignificant over the northern edge of the tropical regions. Overall, the trends show a latitudinal gradient with stronger warming in the SH tropics.

The spatial patterns of the seasonal mean trends are presented in the Appendix A section (Figs. A1-A4), offering a detailed view of how the cold-point temperature trends vary with season. The analysis shows warming across all seasons, with especially pronounced warming over the western hemisphere in boreal winter, the equatorial region during boreal spring, and the southern tropics during boreal autumn. The latitudinal gradient of the temperature trends results from the trend distributions in boreal summer and autumn. Overall, the reanalysis datasets align well with observations in terms of magnitude and spatial patterns of the cold point warming that occurred over the last 20 years.

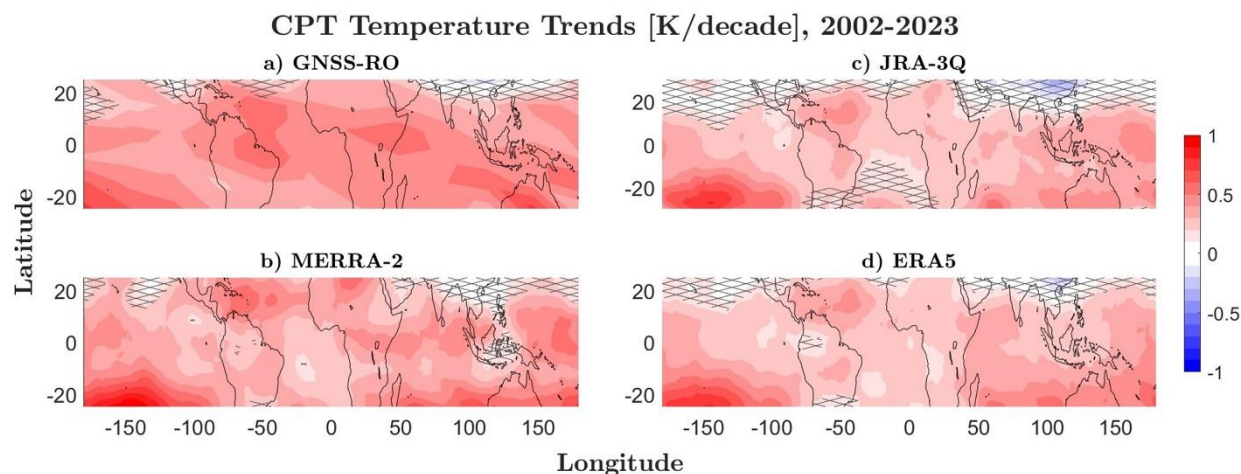



Figure 5. Long-term (2002–2023) temperature trends at the cold-point tropopause (CPT) based on the (a) GNSS-RO, (b) MERRA-22, (C) JRA-3Q, and (d) ERA5 datasets. Trends not significant at the 95% confidence level are marked with grey cross-hatching.

The investigation of the seasonal cycle and the vertical structure of TTL temperature trends over 30°N–30°S highlights discernible patterns across the satellite and reanalysis datasets (Fig. 6). Overall, all datasets exhibit approximately the same behavior. Data from observations indicate substantial warming exclusively between July and December, occurring only at altitudes lower than 150 hPa. In contrast, reanalysis data highlight a broader upper-tropospheric warming trend that extends into the lower stratosphere, reaching up to 80 hPa during boreal spring and summer, 50 hPa in boreal winter, and as high as 30 hPa in October. Conversely, stratospheric cooling is strongest during boreal summer in all data sets. In the upper troposphere, the seasonal cycle of the temperature trends is nearly opposite to that of the lower stratosphere, with the largest warming occurring during boreal late summer and fall. An exception to these opposite seasonal cycles is the October warming which is evident across all altitudes.

Temperature trends at the cold point show a clear seasonal cycle, with two pronounced features. First, warming maximises during boreal winter and spring (Zolghadrshojaee et al., 2024), which is especially evident in the MERRA-2 and ERA5 datasets. This strongest cold point warming during boreal winter could be related to dynamical changes such as a pronounced weakening of the BDC during this time of year. In addition to the boreal winter, the cold point shows a strong warming signal in October, similar to the lower stratosphere, in all data sets. This warming signal might also be related to the weakening of the BDC, as it coincides with a strong negative temperature trend in the Southern Hemisphere polar regions during this month (Ladstädter et al., 2023). Note that the monthly cold point trends are not significant in most datasets. Overall, the reanalyses reproduce the observed temperature trends very well in terms of their seasonal and vertical structure. The best agreement is found for JRA-3Q, with the other data sets showing slightly more pronounced warming trends in the upper TTL and lower stratosphere.

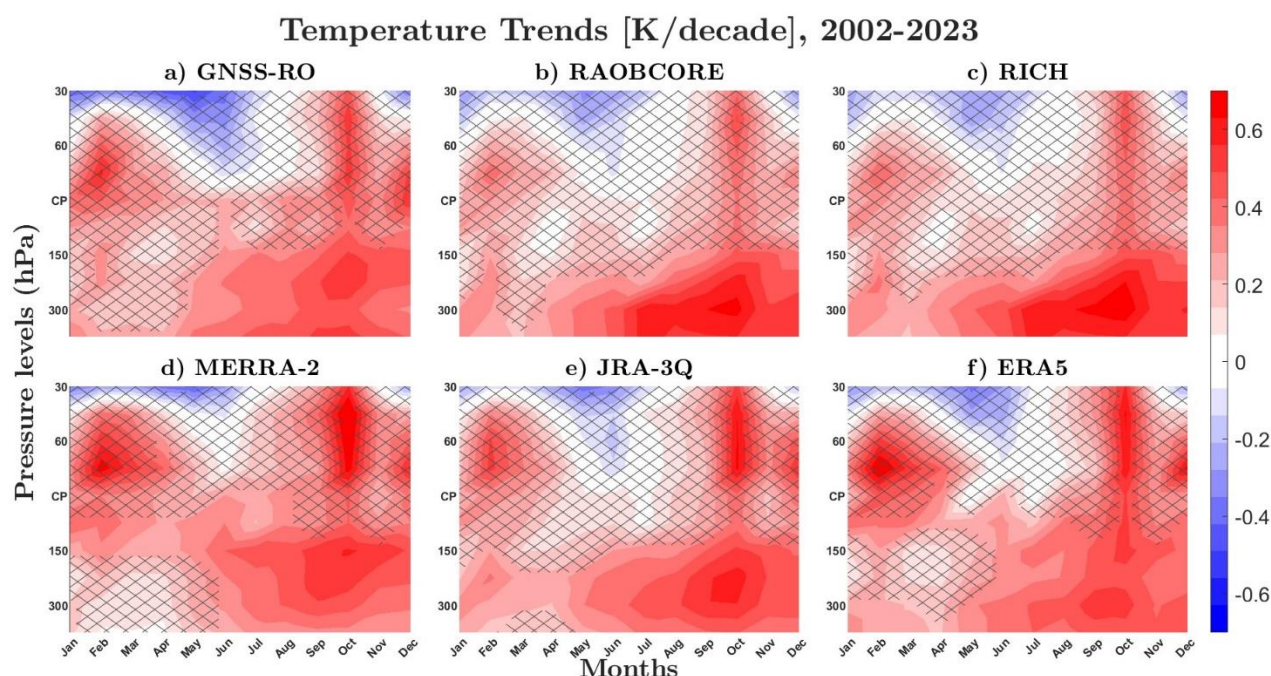



Figure 6. The seasonal cycle of long-term temperature trends (2002-2023) at the cold point (CP) and different pressure levels based on the (a) GNSS-RO, (b) RAOBCORE, (C) RICH, (d) MERRA-2, (e) JRA-3Q, and (f) ERA5 datasets (30°N–30°S). Trends not significant at the 95% confidence level are marked with crosses.

The spatial patterns of temperature trends at the cold point are known to be correlated with temperature trends in the lower stratosphere and anticorrelated with temperature trends in the upper troposphere (e.g., Lin and Emanuel, 2024; Zolghadrshojaee et al., 2024). Here we tested if the reanalysis data sets show this well-known characteristic by calculating the Pearson correlation coefficients between the gridded trends at the cold point and at all other levels (Fig. 7). The CPT temperature trend patterns exhibit striking similarities with temperature trend patterns at levels slightly below (100 hPa) and above (70 hPa), particularly during the boreal winter and fall, for all data sets. These correlations can extend up to 60 and even 50 hPa, depending on season and data set, with the strongest coupling to higher levels found for ERA5.

Local tropospheric temperature trends are anti-correlated with cold-point (and therefore lower stratospheric) temperature trends via regional dynamical or radiative processes (e.g., Kuang and Bretherton, 2004; Lin and Emanuel, 2024). Significant negative correlations are observed consistently at the 300 hPa and 400 hPa levels with the exception of the boreal spring season. GNSS-RO, JRA-3Q, and ERA5 datasets exhibit stronger negative correlations compared to MERRA-2 most of the time. There is also some seasonal variability, with stronger negative correlations during boreal winter and summer. Despite

some differences among the datasets, the overall patterns remain consistent, highlighting a robust vertical structure of temperature trends in the upper troposphere and lower stratosphere.

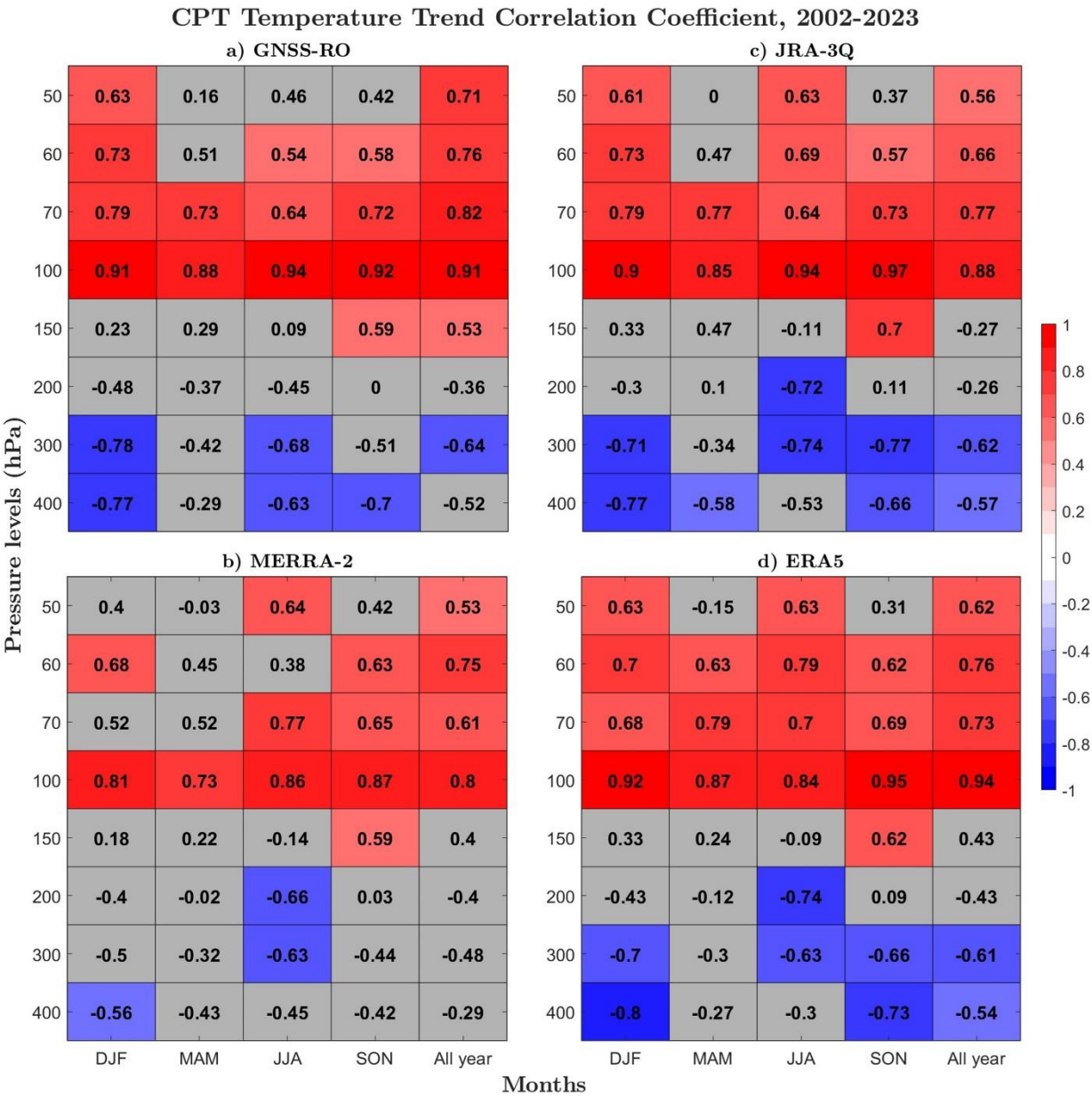



Figure 7. Correlation coefficients between gridded temperature trends at the cold-point tropopause (CPT) and the different pressure levels as given on the y-axis for (a) GNSS-RO, (b) MERRA-2, (C) JRA-3Q, and (d) ERA5 for 2002–2023 (30°N–30°S). Correlation coefficients in grey boxes are not significant at the 95% confidence level.

5. Temperature trends for 1980 to 2001

The tropical cold point tropopause experienced significant cooling before the year 2000 (e.g., Zhou et al., 2001; Seidel et al., 2001; Wang et al., 2012) and here we evaluated if the reanalysis data sets reproduce the magnitude and seasonal variations of

340   these cooling trends. Figure 8 illustrates the tropical average temperature trends for 1980-2001 based on radiosonde data and reanalysis. Unadjusted and adjusted radiosonde data show increasing cooling with altitudes between 100 and 70 hPa, with nearly all trend estimates being significant. The Nearby Level approach results in cooling trends ranging from -0.5 ± 0.31 K per decade to -0.30 ± 0.46 K per decade with the 100 hPa based trend estimates being significant while the 70 hPa based estimates show slightly lower values and larger error bars resulting in non-significant trends (Fig. 8b). Reanalyses also exhibit

345   cold point cooling, though the trends are notably weaker compared to observations, particularly in the case of MERRA-2.

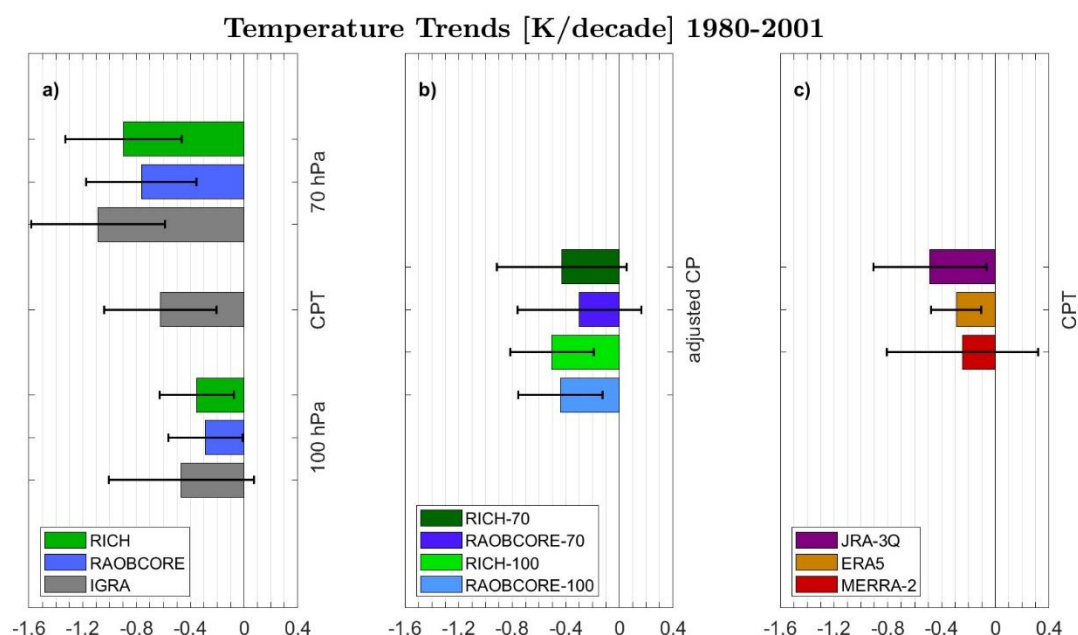



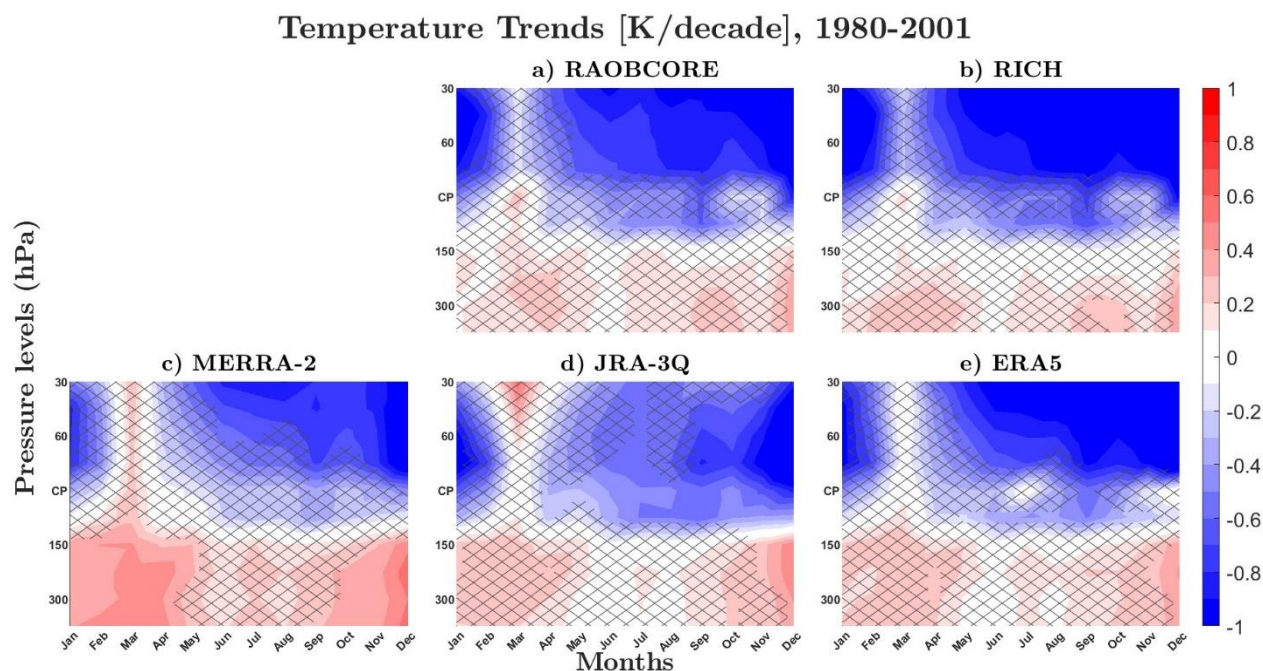
Figure 8. Tropical average (30°S-30°N) temperature trends during 1980-2001 for (a) the cold-point tropopause (CPT) and nearby pressure levels for unadjusted (IGRA) and adjusted (RAOBCORE, RICH) data; and (b) the CPT trends based on adjusted data
350   **and the Nearby Level approach and (c) the CPT trends based on reanalysis data. Error bars show the ± 2 sigma confidence intervals of the trend taking into account adjusted effective sample size.**

Figure 9 illustrates the seasonal cycle and vertical structure of TTL temperature trends over the 30°N–30°S region, based on RAOBCORE, RICH, MERRA-2, JRA-3Q, and ERA5 data for the period 1980–2001. Comparing the adjusted radiosonde

355   trends to the same analysis for 2002-2023 (Fig. 6) reveals smaller warming rates in the upper troposphere (300-400 hPa) for the earlier period reaching only values of up to 0.3 K per decade instead of 0.6 K per decade. This feature of temperature trends being less positive/more negative for the 1980-2001 time period is found at all levels with the differences increasing above the

level of 150 hPa. In the lower stratosphere the differences maximize with cooling trends of around -0.75 K per decade for 1980-2001 compared to a warming trend of around 0.2 K per decade for 2002-2023.

360   The seasonal cycle of the temperature trends shows that the 1980-2001 lower stratospheric cooling is only interrupted in March when temperature trends are around zero for the radiosonde data. The absence of cooling in March might be related to the weakening of the BDC in this month, as the Arctic polar vortex has intensified during NH spring over 1979-1999 (Langematz and Kunze, 2006). This signature of the BDC trends in the seasonal cycle of the tropical cold point temperature trends is similar to the strong October warming found for 2002-2023.

365   All three reanalyses reproduce the seasonal and height variations of the observed temperature trends. Upper tropospheric warming is overestimated by MERRA-2 and also somewhat by JRA-3Q. Similarly, both data sets show slightly weaker cooling at the stratospheric levels. ERA5 temperature trends are most similar to the RAOBCORE and RICH temperature trends.

370   **Figure 9. The seasonal cycle of long-term temperature trends (1980-2001) at the cold point (CP) and different pressure levels based on the (a) RAOBCORE, (b) RICH, (c) MERRA-2, (d) JRA-3Q, and (e) ERA5 datasets (30°N–30°S). Trends not significant at the 95% confidence level are marked with crosses.**

6. Discussion and Summary

375   Results are summarized in Fig. 10, which shows the linear trends in tropical average temperature at the CPT across various observational and reanalysis datasets for the periods 1980-2001, 2002-2023, and 1980-2023. The extended period from 1980-

2023 displays general cooling trends of up to -0.33 ± 0.12 K per decade (based on JRA-3Q data). While the 1980-2023 cooling is less pronounced than cooling before 2000 (for the 1980-2001 period), the uncertainties are smaller resulting in significant trends for the observational data sets. Given that the radiosonde observations underestimate the warming trends for the last two decades when compared to GNSS-RO, the cooling trends for the full time period might be closer to zero than implied by the radiosonde time series. Cooling in JRA-3Q is larger than in observations, while MERRA-2 and ERA5 have slightly smaller and not significant cooling rates.

Warming trends over 2002-2023 are only significant for GNSS-RO satellite observations and the reanalysis data. The latter reproduces not only the magnitude of the warming trends but also the spatial distributions and seasonal variations. Similar to the observations the reanalysis tropopause trends display anticorrelations with patterns of tropospheric temperature trends, and regions of strongest cold-point warming are found to show slight cooling trends in the upper troposphere.

Overall, the comparison in Fig. 10 highlights a significant shift in temperature trends from cooling in the earlier period (1980-2001) to warming in the later period (2002-2023) across all pressure levels and datasets, indicating a robust change in the tropical tropopause temperature dynamics post-2002 (Tao et al., 2023). These changes are consistent with temperature trend changes in the lower stratosphere, where pronounced cooling trends observed before 2000 disappeared and turned into warming trends after 2000. Changes in the Brewer-Dobson circulation in response to changing ozone depleting substances and increasing GHGs have been identified as important drivers of global and tropical lower stratospheric cooling (e.g., Arblaster et al., 2014). Model results suggest that lower stratosphere and cold point tropopause cooling before 2000 is caused by enhanced upwelling and decreases in tropical ozone (e.g., Polvani et al., 2017).

Here we examined if the reanalysis trends in tropical upwelling are consistent with trends in cold point temperature over the three time periods. Reanalysis estimates of vertical velocities are known to suffer from numerical noise as they are computed indirectly from horizontal divergence. In consequence, even relatively small assimilation increments applied to the horizontal winds can have a large influence on the vertical velocity estimates (e.g., Uma et al., 2021). However, since no direct observations of tropical upwelling exist, reanalysis estimates of the vertical velocities have been used in many studies.

Figure 10 shows that upwelling from all three reanalyses have opposite trends for 1980-2001 compared to 2002-2023 consistent with the cold point and lower stratosphere temperature trends. While the vertical residual circulation increased before 2000 consistent with cold point cooling, the circulation trends turned to zero (MERRA-2, JRA3Q) or became positive (ERA5) after 2000 consistent with cold point warming. Note that negative values of the vertical velocity given in Pa s^{-1} correspond to upwelling. The circulation trends over the full time period (1980-2023) suggest increased upwelling (MERRA-2, JRA-3Q) or zero trends (ERA5). This is in very good agreement with the results from Abalos et al. (2015), who compared three different estimates of upwelling for each reanalysis. Overall, the magnitude of the upwelling trends in Fig. 10 shows large differences between the reanalyses, with MERRA-2 being the most negative and ERA5 being the most positive.

However, the change of the trend between the time periods is very similar for all data sets and consistent with the shift in temperature trends.

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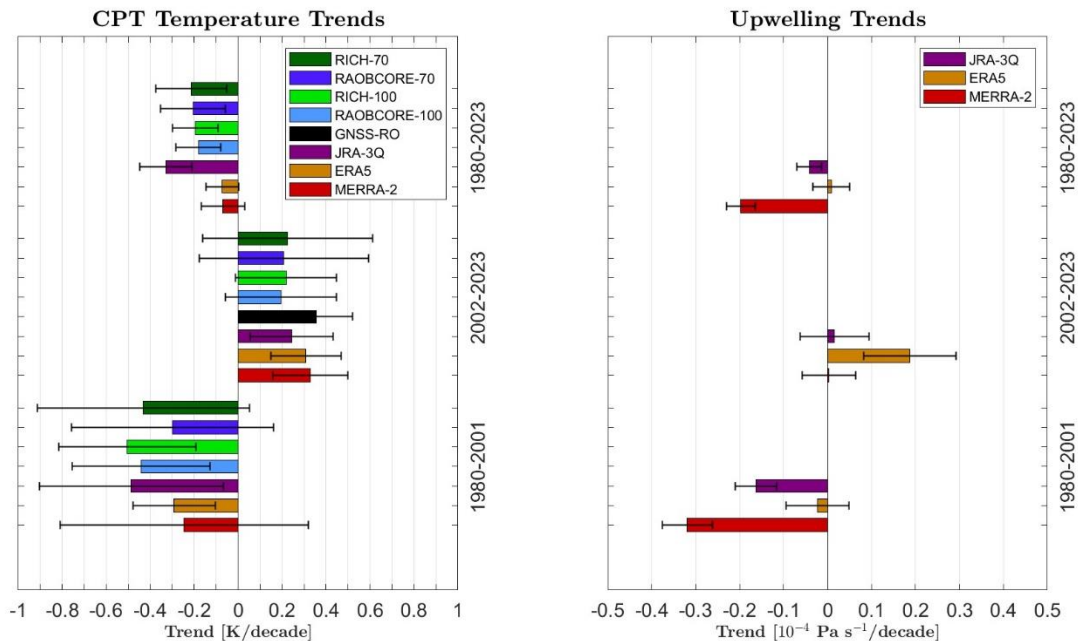
Figure 10. Tropical average (30°S-30°N) of trends in cold-point tropopause (CPT) temperature and tropical upwelling based on different observational and reanalysis datasets for three periods: 1980-2001, 2002-2023, and 1980-2023. The reanalysis tropical upwelling is the vertical residual circulation averaged over 100 and 70 hPa. Error bars show the ± 2 sigma confidence intervals of
415 **the trend taking into account adjusted effective sample size.**

In a last step, we compared the seasonal cycle of the cold point trends with the seasonal cycle of upwelling trends (Fig. 11). If the changes in the BDC are driving the temperature trends, one can expect trends in both variables to have the same seasonality. Cold point temperature trends from GNSS-RO show a clear seasonal cycle with stronger warming during NH winter and in October. The reanalyses mostly reproduce this seasonal cycle with ERA5 overestimating the amplitude, while MERRA-2
420 shows a weaker seasonal signal.

The comparison of the seasonal cycle in temperature and upwelling trends shows good agreement for ERA5 and JRA-3Q. For both data sets, strong decrease in upwelling during NH winter as well as in September and October occur at the same time of year as strong cold point warming. Therefore, these two signals are consistent with current discussions that suggest changes in upwelling to be an important driver of TTL temperature changes (e.g., Polvani et al., 2017). While the seasonality of
425 upwelling and temperature trends in ERA5 and JRA-3Q agree for most of the year, the months of November and December are a bit of an exception to this pattern as upwelling increases or is close to zero and the temperature continues to show positive trends. Furthermore, a spike in the decrease in upwelling in June somewhat obscures the seasonal signal in the upwelling

trends. Finally, upwelling and temperature trends in MERRA-2 do not share a common seasonality with the temperature trend having a relatively flat seasonal cycle while the seasonal cycle of the upwelling trend is similar to the one found in ERA5 and 430 JRA-3Q.
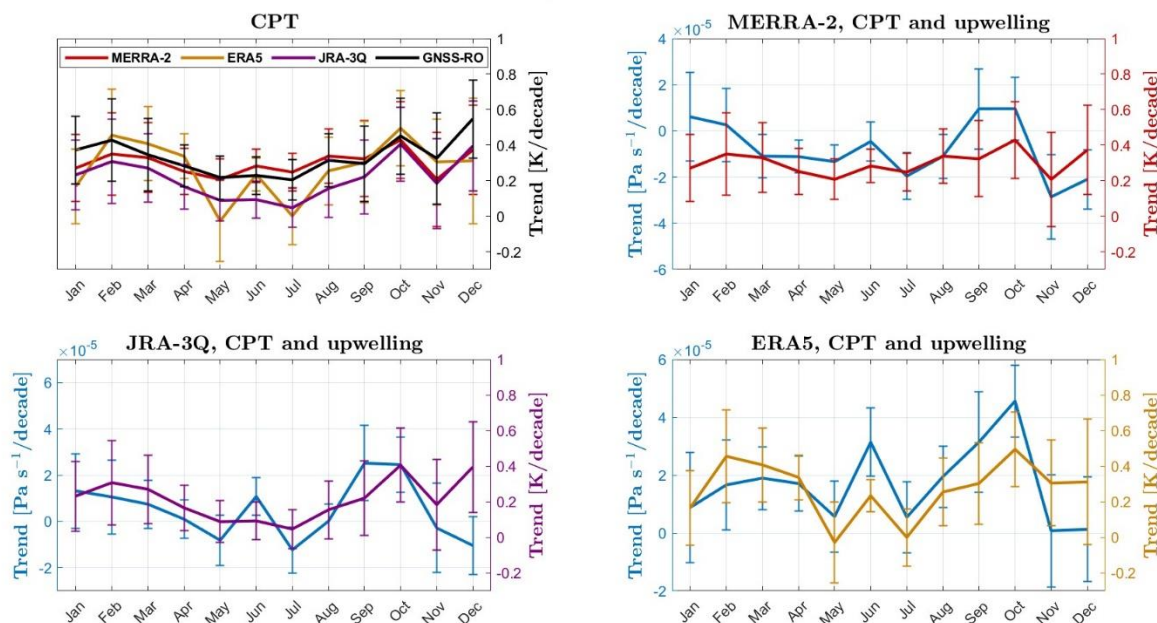
Figure 11. Seasonal cycle of tropical trends average (30°S-30°N) in cold-point tropopause (CPT) temperature and tropical upwelling based on different observational and reanalysis datasets for 2002-2023. The reanalysis tropical upwelling is the vertical residual circulation averaged over 100 and 70 hPa. Error bars show the ± 2 sigma confidence intervals of the trend taking into account 435 adjusted effective sample size.

Conclusion

A significant shift in tropical cold point temperature trends from cooling in 1980-2001 to warming in 2002-2023 is found in observations. Reanalysis data sets reproduce the robust change in the tropical tropopause temperature trends. The trends in upwelling from reanalyses differ significantly from each other. However, these upwelling trends are consistent with the 440 changes of cold point and lower stratosphere temperatures and show a shift from increased upwelling before 2001 to decreased or zero upwelling after 2001. This implies that while the absolute vales of reanalyses upwelling trends remain poorly constrained, their decadal and multidecadal variability is dynamically consistent with changes in temperature.

The here identified changes highlight a strong decadal to multidecadal variability of TTL temperatures which could be related 445 to internal variability of the climate system. Increasing greenhouse gases in the troposphere and stratosphere, changing ozone trends and variations in BDC upwelling can all play a role for TTL temperature changes. Since cold point temperature trends

largely control the amount of water vapor entering the stratosphere, they indirectly influence stratospheric chemistry, radiation and surface climate. CMIP6 models are known to have difficulties in the TTL region and a poor representations of stratospheric water vapor (e.g., Keeble et al., 2021). Quantifying and understanding the decadal variability and long-term TTL temperature

450 trends found in observations can help to evaluate and improve the representation of the stratosphere in models.

Data Availability: Unadjusted IGRA radiosonde data are accessible via the NOAA Integrated Global Radiosonde Archive website at https://www.ncei.noaa.gov/products/weather-balloon/integrated-global-radiosonde-archive (last accessed: January 2025; Durre et al., 2006). Adjusted radiosonde data can be obtained from the University of Vienna RAOBCORE website at

455 https://imgw.univie.ac.at/forschung/klimadiagnose/raobcore/ (last accessed: January 2025; Haimberger, 2007; Haimberger, 2008). GNSS-RO data are available through the COSMIC Data Processing Center at https://www.cosmic.ucar.edu/what-we-do/data-processing-center/data (last accessed: January 2025). For the reanalysis data, we utilized monthly averaged MERRA-2 temperature fields on pressure levels (Global Modeling and Assimilation Office, 2015). Additionally, we incorporated monthly mean ERA5 temperature data (Copernicus Climate Change Service, Climate Data Store, 2023) and JRA-3Q monthly

460 mean data (Numerical Prediction Division, Information Infrastructure Department, 2022).

Author contribution: MZ performed the analysis and wrote the manuscript. ST developed the research question and guided the research process. RPK provided the GNSS-RO data. SMD provided the reanalysis data. LH provided ROABCORE and RICH data. All authors took part in the process of manuscript preparation.

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Competing interests: The contact author has declared that none of the authors has any competing interests.

Financial support: This research has been supported by the Natural Sciences and Engineering Research Council of Canada (NSERC, grant no. RGPIN-2020-06292).

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Appendix A

635
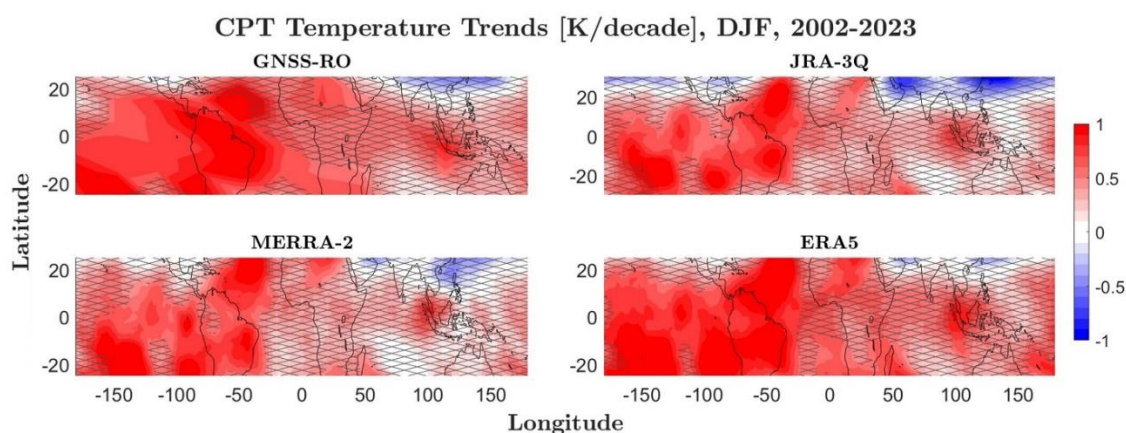
Figure A1. Same as the Figure 5 but for DJF.
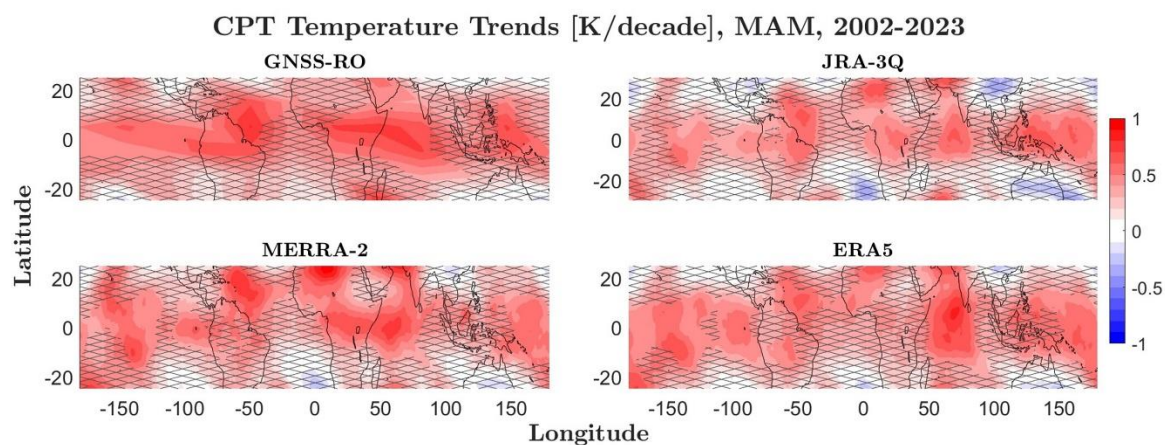
Figure A2. Same as the Figure 5 but for MAM.

640
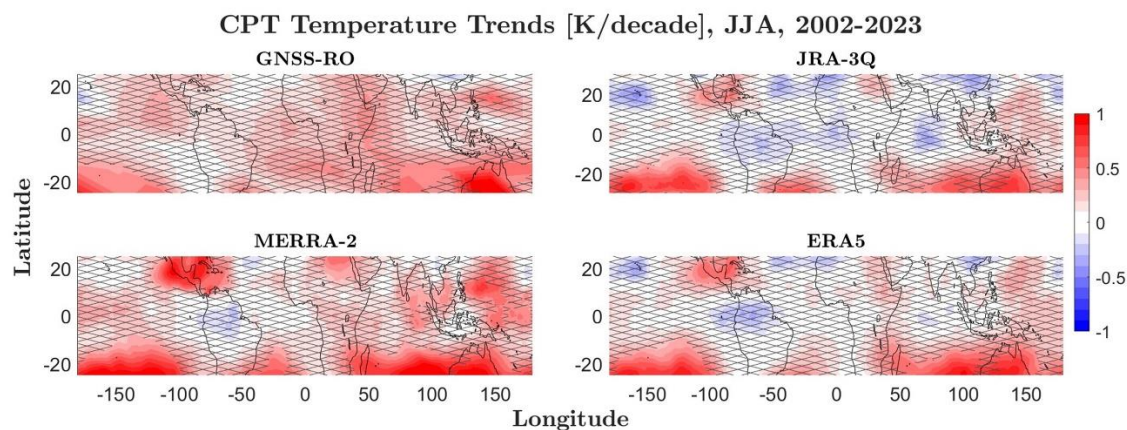
Figure A3. Same as the Figure 5 but for JJA.

645
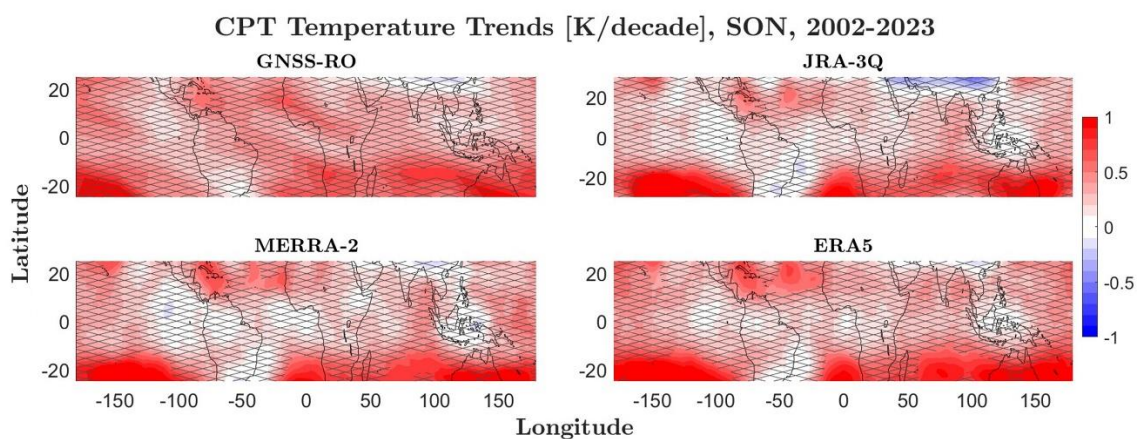
Figure A4. Same as the Figure 5 but for SON.