



Narrowing of Lagrangian Cold-Point Pathways into the Tropical Stratosphere

Mona Zolghadrshojaee¹, Sean M. Davis², Kirstin Krüger³, and Susann Tegtmeier¹

¹Institute of Space and Atmospheric Studies, University of Saskatchewan, Saskatoon, Canada

²National Oceanic and Atmospheric Administration Chemical Sciences Laboratory, Boulder, CO 80305, USA

³Meteorology and Oceanography Section, Department of Geosciences, University of Oslo, Oslo, Norway

Correspondence to: Susann Tegtmeier (susann.tegtmeier@usask.ca)

Abstract. The tropical tropopause layer (TTL) plays a crucial role for air mass transport from the well-mixed convective troposphere into the radiatively controlled stratosphere. The minimum temperatures found at the cold-point tropopause are particularly important for regulating water vapour entry into the stratosphere. Extensive research has revealed a slow widening of the tropical belt in the troposphere while the Intertropical Convergence Zone (ITCZ) was found to become narrower. It remains unclear whether such structural changes also occur for the stratospheric entry region near the cold-point tropopause and how this would impact long-term cold-point temperature trends.

Here, we use 44 years (January 1981 - August 2024) of Lagrangian trajectory simulations driven by MERRA-2 reanalysis to quantify long-term changes in the temperature and meridional width of the Lagrangian cold-point (LCP) region. We introduce a trajectory-based metric, defined as the 90th-percentile latitude range of the LCP distribution, to diagnose the spatial extent of stratospheric entry pathways. We find statistically significant cooling of LCP temperatures (-0.098 ± 0.027 K per decade) and a narrowing of the LCP width ($-0.350 \pm 0.098^\circ$ per decade). A decomposition of LCP-weighted temperature fields reveals that both local temperature variability and changes in transport pathways contribute to the observed long-term cooling. Additional analysis of transport width in vertical layers around the LCP (± 15 hPa) reveals a consistent narrowing of tropical upward transport pathways, although this narrowing is weaker than that of the LCP distribution itself. Together, these findings indicate that air parcels entering the tropical lower stratosphere increasingly experience their coldest temperatures within a narrower latitude band.

1 Introduction

The tropical tropopause layer (TTL) is a transitional region between the convectively active troposphere and the radiatively controlled stratosphere (Fueglistaler et al., 2009). Rather than being a sharply defined boundary, the TTL represents a gradual transition zone with complex thermodynamic and dynamical characteristics, extending roughly from 14–19 km altitude



(potential temperature of ~ 355 – 425 K, 150 – 70 hPa). A key feature of this region is the level of zero radiative heating, which marks the shift from net radiative cooling to heating (Highwood and Hoskins, 1998; Folkins et al., 1999). Another important feature of the TTL is the temperature minimum, termed the cold-point tropopause, where air parcels transitioning from the troposphere to the stratosphere encounter final dehydration. Understanding air transport through this region is therefore critical for stratospheric water vapor, which exerts strong radiative and chemical influences on climate.

In a Lagrangian framework, the amount of water vapor entering the stratosphere is controlled primarily by the minimum saturation mixing ratio encountered along an air parcel's trajectory—known as the Lagrangian Dry Point. Given the strong temperature dependence of the saturation vapor pressure in the Clausius-Clapeyron equation, the saturation mixing ratio minimum generally occurs at the coldest temperature experienced by the air parcel (Brewer, 1949; Holton et al., 1995). In consequence, the term Lagrangian Cold Point (LCP) is often used interchangeably with Lagrangian Dry Point, as will be done here. The concept of a Lagrangian cold point controlling stratospheric water vapour has been used in several earlier trajectory and observational studies (e.g., Fueglistaler et al., 2005; Schiller et al., 2009; Wright et al., 2011).

Another process that can impact lower stratospheric water vapor is deep convection depositing ice particles above the cold point tropopause, where their sublimation can cause a moistening effect (e.g., Corti et al., 2008; Steinwagner et al., 2010). While the global impact of convective ice on stratospheric water vapor was found to be small (e.g., Schoeberl et al., 2018), it can influence the interannual variability (Avery et al., 2017) and seasonal cycle (Wang et al., 2019) of water vapor in the tropical lower stratosphere. Deep convection is also known to impact cold-point temperature and height. Pronounced cold-point temperature minima over the equatorial western Pacific, South America, and western Africa broadly coincide with regions of active convection (e.g., Randel et al., 2003; Kim and Son, 2012). In particular, overshooting convection can deepen the convective layer and lower the cold-point temperature (e.g., Randel and Park, 2019; Proud and Bachmeier, 2021). Tropical convection such as the Madden-Julian oscillation (MJO), is known to impact the intraseasonal variability of the cold-point properties (Zhou and Holton, 2002; Randel et al., 2003). On seasonal time scales, the shift of the Intertropical Convergence Zone (ITCZ) can change the latitudinal distribution of cold-point occurrences (Jain et al., 2011), while interannual anomalies in cold-point temperatures are, among other things, associated with the El Niño-Southern Oscillation cycles (e.g., Zhou and Holton, 2002).

Lagrangian trajectory methods provide a powerful framework for studying transport and dehydration in the TTL, since they track the movement of individual air parcels using wind fields from meteorological reanalyses. These methods have been widely used to identify where and how dehydration occurs, as the highest density of LCPs corresponds to the lowest temperatures and the most efficient gateways for air entering the stratosphere (Bonazzola and Haynes, 2004; Fueglistaler et al., 2005; Krüger et al., 2008, 2009). Lagrangian studies have successfully linked seasonal and interannual variability in cold-point temperature to observed variations in stratospheric water vapor (Fueglistaler et al., 2005; Schoeberl et al., 2012; Hegglin et al., 2014; Wang et al., 2015; Avery et al., 2017). In many cases, Lagrangian models treat the dehydration of air entering the stratosphere as an instantaneous and irreversible removal of water vapor from the parcel when a threshold supersaturation



65 value is reached and do not include cloud microphysics. This approach effectively treats cloud formation and ice sedimentation as an instantaneous process that removes water vapor from the parcel.

Eulerian cold-point diagnostics are useful for characterizing large-scale thermal structure in the TTL but may not accurately represent the lowest temperatures sampled along trajectories, nor the specific pathways that govern dehydration. Because Eulerian cold points reflect uniform spatial averages not taking into account individual air parcel pathways and sampling
70 patterns, they can underestimate true cold-point temperature variability and obscure key transport features. In contrast, Lagrangian cold points follow the minimum temperature along parcel paths and therefore provide a more physically grounded measure of where stratospheric entry and dehydration occur (Fueglistaler et al., 2005; Fueglistaler and Haynes, 2005; Krüger et al., 2008).

Variability in the TTL is closely connected to broader structural changes across the tropical atmosphere such as the tropical
75 belt and ITCZ. The tropical belt is typically defined dynamically as the region encompassing both the rising and sinking branches of the Hadley circulation (e.g., Frierson et al., 2007; Johanson and Fu, 2009). The ascending branch near the equator transports heat and moisture upward, driving deep convection and heavy precipitation, while the descending branches at the subtropical margins produce drier and more stable conditions. Numerous metrics have been used to quantify the width of the tropical belt and its long-term changes. These include upper-tropospheric metrics such as the tropopause break, subtropical jet
80 latitudes or meridional ozone gradients (Seidel and Randel, 2007; Manney and Hegglin, 2018; Harzer et al., 2025) and lower-tropospheric metrics, such as the Hadley cell edge, precipitation or surface wind minima in the subtropics (e.g., Lu et al., 2007; Staten et al., 2018). Trends of the tropical width can be impacted by inhomogeneities in older reanalyses, choice of thresholds and internal variability (Birner, 2010; Davis and Rosenlof, 2012; Davis and Davis, 2018; Grise et al., 2019). Studies that accounted for the methodological differences between the metrics found a widening of the tropical belt of about 0.25 to 0.5°
85 latitude per decade (Staten et al., 2018; Staten et al., 2020).

At the same time, structural changes within the deep-convective core of the tropics reveal the opposing width trend. The ITCZ, which marks the ascending branch of the Hadley circulation, has narrowed in recent decades, contrasting with the poleward expansion of the circulation's descending branches. The ITCZ width can be defined using low-level mass convergence, hydrological measures such as rainfall data sets, mid-troposphere mass fluxes and measures of high reaching convection such
90 as outgoing longwave radiation (e.g., Byrne et al., 2018). Observational and reanalysis products based on these measures document a significant contraction of the ITCZ, especially in the Pacific (Wodzicki and Rapp, 2016; Su et al., 2020), which is consistently reproduced by global climate models (Lau and Kim, 2015; Byrne and Schneider, 2016; Su et al., 2017; Donohoe et al., 2019; Schiro et al., 2019).

Despite extensive evidence for ITCZ narrowing and tropical belt widening, it remains unclear how the potential changes in the
95 intensity, height, and frequency of deep convection will affect the cold-point tropopause where dehydration and stratospheric entry take place. Given that the cold-point distribution and variability is strongly coupled to the tropical deep convection, a long-term change of the cold-point distribution consistent with the ITCZ narrowing seems possible. Earlier Lagrangian studies primarily focused on water vapor and temperature variability (e.g., Fueglistaler et al., 2005; Krüger et al., 2008; Schoeberl and



Dessler, 2011), rather than on the meridional extent of dehydration pathways. Fueglistaler and Haynes (2005) demonstrated that the latitudinal density distribution of LCPs varies with large-scale circulation anomalies, and Hu and Fu (2007) noted that periods of broader LCP spreading may indicate changes in tropical upwelling. These findings were qualitative and event-based, and to our knowledge, no study has systematically quantified long-term trends in the width of the LCP region, nor examined whether changes in transport pathways near the cold-point tropopause and analysed how this might be related to changes in cold-point temperature and stratospheric water vapour. This gap limits current understanding of the vertical structure of tropical circulation and its influence on TTL dehydration and lower-stratospheric water vapor.

We address this gap by introducing a trajectory-based dynamical metric that quantifies the meridional width of the LCP region. Using a 44-year global archive of forward trajectories driven by MERRA-2 winds, we assess long-term changes in both LCP temperature and LCP width, analyse the vertical structure of transport surrounding the LCP, and separate the relative contributions of temperature variability and evolving pathway geometry. This approach allows us to evaluate whether and how stratospheric entry pathways changed over recent decades and how these structural changes relate to TTL temperature changes. Section 2 describes the trajectory dataset and Eulerian cold-point temperatures. Section 3 outlines the methods used to identify Lagrangian cold points, quantify their meridional width, and diagnose long-term trends. Sections 4 and 5 present the key results and discuss their implications for tropical widening and stratospheric dehydration. Section 6 summarizes the main findings and discusses their implications for TTL tropical width changes.

2 Data

Two complementary sampling approaches are used to characterize cold-point temperatures and their spatial distribution in the TTL. In the Eulerian framework, the coldest temperatures are diagnosed from gridded atmospheric (reanalysis) temperature profiles, while the Lagrangian framework identifies the lowest temperature encountered along air parcel trajectories.

2.1 Eulerian cold-point temperature

Eulerian cold-point (ECP) temperatures are derived from the Modern-Era Retrospective Analysis for Research and Applications, Version 2 (MERRA-2; Gelaro et al., 2017), which was produced with an upgraded version of the Goddard Earth Observing System Model, Version 5 (GEOS-5) data assimilation system. MERRA-2 provides a realistic representation of the major characteristics of the temperature structure within the TTL with a warm bias in cold-point temperature similar to other reanalysis data sets (Tegtmeier et al., 2020; SPARC, 2022). Interannual variability and long-term trends of MERRA-2 TTL temperatures are consistent with observational data sets.

Six-hourly instantaneous MERRA-2 temperature profiles are available on a $0.625^\circ \times 0.5^\circ$ longitude-by-latitude grid on 72 model layers for 1980–2024 (GMAO, 2015a). The six-hourly profiles are used to extract the minimum temperature in each vertical column within 15°S – 15°N . The 15°S – 15°N band is chosen to encompass the deep tropics. In a final step, the minimum temperatures are averaged to derive the monthly mean, tropical mean Eulerian cold-point temperature fields. These fields



130 represent the background lowest temperatures resolved on the MERRA-2 reanalysis grid and are used for comparison with
LCP-sampled temperatures.

2.2 Lagrangian cold-point temperature

Lagrangian cold points are diagnosed from a multi-decadal global trajectory dataset generated using a trajectory model driven
by 3-hourly MERRA-2 wind and temperature fields (GMAO, 2015b, c, d). The simulation initializes approximately 10,000
135 air parcels every 14 days from 1980 to 2024 on the 370 K isentropic surface, distributed evenly between 60°S and 60°N. For
the tropics, the 370 K level lies within the upper TTL (~ 15–16 km), where air parcels begin their final ascent toward the
stratosphere, ensuring that the trajectories adequately sample the cold-point region relevant for dehydration. Previous trajectory
studies have shown that isentropic levels near 360–380 K provide good coverage of air parcels that subsequently cross the cold
point and enter the lower stratosphere (e.g., Fueglistaler et al., 2005; Krüger et al., 2008, 2009), and the choice of 370 K is
140 consistent with this framework and above MERRA-2's level of zero radiative heating. Trajectories are integrated forward
during their transport through their stratosphere until they descend below 400 hPa. In addition to a realistic representation of
the TTL temperature structure, the frequency distribution of the MERRA-2 cold-point pattern agrees well with other reanalyses
(SPARC, 2022). Furthermore, the absolute values and interannual variability in MERRA-2 tropical upwelling, which drives
the upward transport of the trajectories, is consistent with upwelling in other modern reanalysis and Chemistry Climate Models
145 (SPARC, 2022).

For each parcel, the LCP temperature is defined as the minimum temperature encountered along the trajectory. The model also
records the parcel location at the LCP (latitude, longitude, pressure, potential temperature). Although the trajectory simulation
includes a basic dehydration parameterization, in which parcels are initialized with 200 ppmv water vapor and supersaturation
greater than 1.2 is removed, the present study focuses solely on temperatures and the spatial distribution of LCPs, rather than
150 on water vapor evolution. We expect the location of the Lagrangian dry point and the LCP to be essentially the same given the
strong dependence of the dry point on temperature.

Cold-point and trajectory information is archived in monthly snapshot files with unique parcel identifiers, enabling
reconstruction of parcel pathways and association of each LCP with its geographic location. This dataset provides a physically
consistent basis for analysing the long-term behaviour and meridional extent of stratospheric entry pathways.

155 3 Methods

3.1 Identification of Lagrangian Cold Points (LCPs)

Air in the tropical upper troposphere and lower stratosphere transitions gradually from tropospheric to stratospheric
characteristics within roughly 14–19 km altitude, equivalent to ~355–425 K potential temperature and 150–70 hPa pressure
(Tuck et al., 1997; Folkins et al., 1999; Marcy et al., 2007; Fueglistaler et al., 2009). The coldest temperatures experienced by
160 ascending air parcels—and therefore the primary dehydration region—occur in the upper part of this transition layer around

90 hPa. As we aim to analyse stratospheric entry characteristics, we focus our analysis on air parcels that reach the 400–420 K potential temperature range between 30°S and 30°N latitude, ensuring that we only consider air parcels that have successfully entered the tropical lower stratosphere.

We use monthly instantaneous trajectory snapshot files from the forward-integrated Lagrangian simulation (Section 2). Each file contains the current position and trajectory history of every air parcel, including the minimum temperature encountered since initialization. For each month from January 1981 through August 2024, we extract all trajectories that fulfil the following two conditions (Fig. 1). First, the trajectories are located within the tropical lower stratosphere (30°S–30°N, potential temperature 400–420 K), and second, the trajectories experienced their cold point no more than 7 months prior to the snapshot time. The 7-month limit is chosen to encompass typical transit times from the cold point into the lower stratosphere while excluding older air masses that have resided in the stratosphere for many months, ensuring that only air parcels that recently passed through the TTL cold point are included.

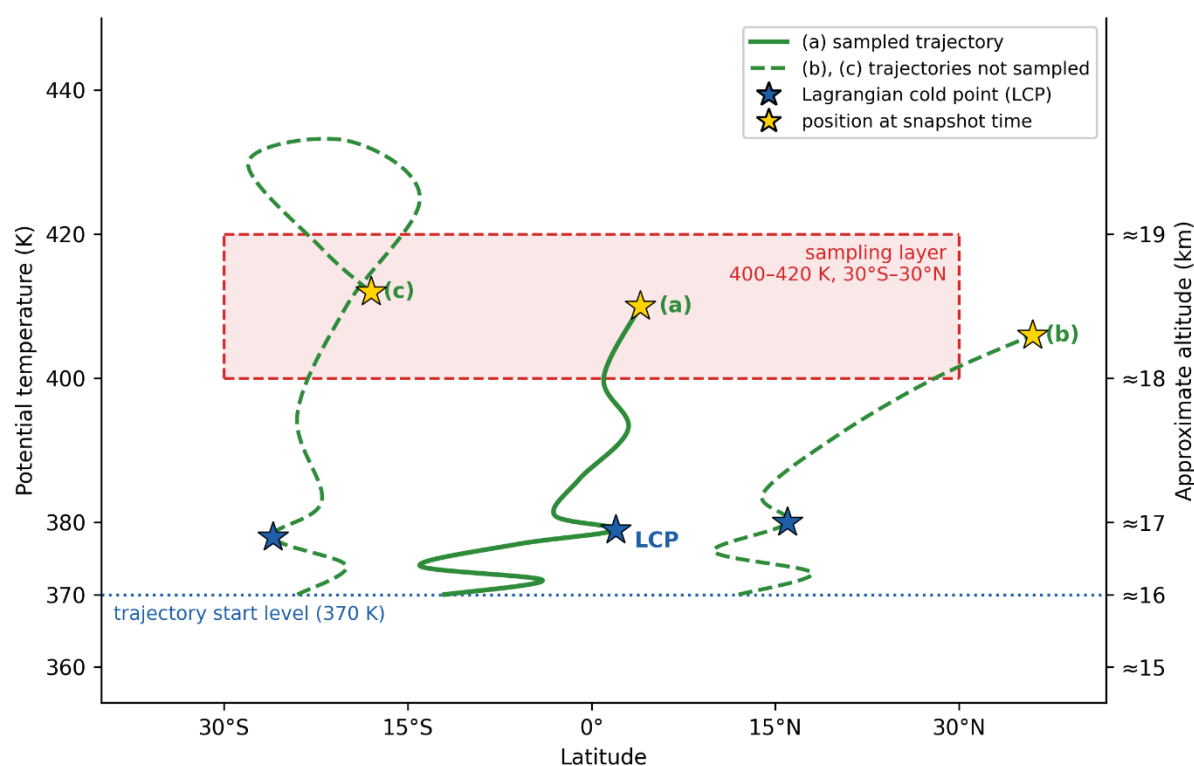


Figure 1. Schematic of the trajectory sampling criteria. Blue stars mark the Lagrangian cold point (LCP); yellow stars mark parcel positions at snapshot time. Trajectory (a) is sampled: it lies in the 400–420 K layer within 30°S–30°N (red box). Trajectories (b) and (c) are not sampled. Altitudes are approximate.



Trajectories already sampled in the most recent month are removed from subsequent months to avoid duplication. Iterating backward through all monthly files yields a unique record for each air parcel that ascended through the tropical cold point during the 44-year period. For each LCP, we retain temperature, latitude, longitude, pressure, potential temperature, time stamp, and parent trajectory identifier. These records form the monthly LCP database used to diagnose climatology, variability, and long-term trends in LCP location and dehydration pathways.

3.2 Trend calculations

To calculate the meridional width of the Lagrangian cold point, we follow the method proposed by Mohr (2013) and first analyse the zonally averaged distribution of the cold points as a function of latitude based on monthly, seasonal, or annual mean fields. In a second step, we calculated for each monthly, seasonal, and annual mean field, the 95th and 5th percentiles of the LCPs latitude distribution. This gives the northern and southern latitude edges so that 90% of all LCPs fall within this range, and the remaining 10% lie outside this range at higher absolute latitudes in each hemisphere. The meridional width of the LCP region based on the 90th percentile latitudinal range is given by Eq. (1) as:

$$W(t) = \phi_{95}(t) - \phi_5(t), \quad (1)$$

where ϕ_5 and ϕ_{95} correspond to the 5th and 95th percentiles of the LCP latitude distribution and t corresponds to time according to the monthly, seasonal, or annual resolution. This 90th percentile latitudinal range of the LCPs is referred to as the latitudinal LCP width. To evaluate the robustness of this percentile-based method, we conducted a sensitivity study by varying the percentile range from 50 to 100 in 1-percentile steps for the calculation of the LCP width trends. For the full 1980-2024 time series, the trend changes by less than 20% for percentiles between 50 and 90, and by less than 10% for percentiles between 80 and 90, ensuring that we are using a stable threshold for estimating trends in LCP width.

Furthermore, monthly LCP temperature and width anomalies are calculated relative to their respective monthly climatologies. Trends are estimated using ordinary least squares regression according to Eq. (2):

$$X(t) = A + B \cdot t + \epsilon(t), \quad (2)$$

Here $X(t)$ represents either temperature or width anomalies, A is the intercept, B is the trend coefficient, t denotes time, and $\epsilon(t)$ are the residuals. All anomaly time series shown in the figures are smoothed with an 8-month running-mean filter for visualization purposes, while trend estimates are computed from the unfiltered series. The uncertainty in of the long-term trends are calculated as the standard error of the slope with the effective sample size adjusted to account for the corresponding lag-1 autocorrelation coefficient (Santer et al., 2000). Significance was tested based on a two-tailed test with a 95 % confidence interval.

To quantify the spatial distribution of transport pathways, the number of trajectories associated with parcels within the 400–420 K layer in a given month is calculated for each $5^\circ \times 5^\circ$ grid cell. The trajectory density (ν) is defined according to Eq. (3):

$$\nu = \frac{n_b}{n_{T.a}} \times 100 \times 10^6, \quad (3)$$



Here n_b is the number of trajectories within a grid cell, n_T is the total number of trajectories, and a is the area of each grid cell (km²), and the resulting trajectory density is in percentage of trajectories per 1 million km².

3.3 Identifying drivers of LCP trends

We aim to isolate the contribution of background temperature changes to long-term LCP cooling from the impact of evolving transport pathways. We, therefore, construct a scenario where the cold point sampling is constant while the background temperature fields change according to their temporal evolution as given by the reanalysis. In the first step, we calculate the monthly mean climatological LCP trajectory-density maps (v_i) on a 5°×5° latitude-longitude grid based on the trajectory cold-point statistics of the full time series. In a second step, we use these density maps to weigh the time-varying Eulerian cold-point temperature fields $T_i(t)$, which are also taken on a 5°×5° latitude-longitude grid (Eq. 4).

$$T_{varT}(t) = \frac{\sum_i v_i(m(t)) \cdot T_i(t)}{\sum_i v_i(m(t))}, \quad (4)$$

where t is the monthly time index in the full record (Jan1981–Aug2024), $m(t)$ is the calendar month (1–12) associated with time t , and i indexes the 5°×5° spatial grid cells.

This approach provides an estimate of the impact of the background temperature variations on the changes in the LCP temperature as the sampling is kept constant. While the scenario does not match a real physical situation, since the temperature and transport fields do not necessarily match each other, it helps to understand what effect temperature variations alone could have.

Another goal of this study is to investigate why the LCP meridional width is changing. A change of the LCP width may arise from changes in the thermal structure of the TTL, from modifications in how parcels are transported into the coldest regions or from a combination of the two effects. We want to analyse whether the LCP width change can also be found in the width of the branch of upward-moving air masses in the upper TTL. To isolate the transport contribution, we examine how the meridional spread of trajectories evolves at different vertical levels relative to the cold point.

In a first step, we determine the monthly tropical mean climatological pressure and potential temperature levels of the LCP. Based on the monthly mean LCP level, we define six consecutive layers of 5 hPa depth, with three layers above and three layers below the LCP. For all trajectories that contribute to the Lagrangian cold point diagnostic, we track their parcel positions while they are transported upwards through these consecutive layers. The latitudinal positions of all trajectories within a given layer form the monthly mean meridional transport distribution. The transport width at each layer is defined as the smallest latitude interval containing 90th of all trajectories within that layer. This percentile-based width quantifies the meridional confinement of air parcels as they approach and depart the cold point, allowing us to assess whether transport pathways throughout the upper TTL have changed over time.

4 LCP climatological distribution and seasonal cycle

The climatologies of LCP temperature and mean occurrence are shown in Fig. 2. LCP temperatures (color shading) are given as a scatter plot, while the LCP spatial density (white contour lines) is sampled on a $5^\circ \times 5^\circ$ latitude-longitude grid. The long-term annual means reveal an extensive region around the equator with LCP temperatures below 190 K, except for the eastern Pacific and the Atlantic regions. Lowest temperatures occur over the maritime continent, with minimum LCP temperatures of approximately 182 K. Consistent with previous findings, regions with the highest LCP density generally align with the lowest temperatures, indicating a strong association between LCP occurrence and thermal conditions (Holton and Gettelman, 2001; Fueglistaler and Haynes, 2005; Fueglistaler et al., 2005; Krüger et al., 2008; Schoeberl and Dessler, 2011). The maximum LCP density reaches up to 1% of trajectories per million km^2 over the coldest regions, suggesting that a relatively large fraction of all air masses experience their final dehydration within these extremely cold areas.

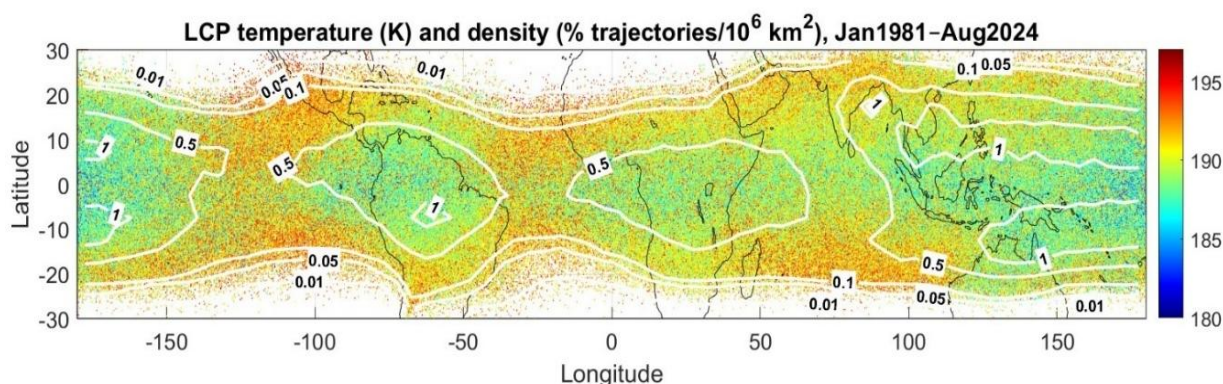


Figure 2. Annual mean climatology of the Lagrangian Cold Point (LCP) temperature (K) (January 1981 to August 2024). White contour lines display the density of LCP occurrence, expressed as the percentage of trajectories per 1 million km^2 .

Figure 3 displays the seasonal climatology of LCP temperature and density for the period January 1981-August 2024. The seasonal distributions exhibit a pronounced annual cycle in both temperature and LCP occurrence. Consistent with the seasonal cycle in background temperature, the overall coldest LCP is found during boreal winter, while much warmer LCPs dominate during boreal summer. The latitudinal distribution also shifts with the seasons from higher-density areas in the northern tropics during boreal summer to the high-density areas in the southern tropics during austral summer. This is consistent with the seasonal fluctuations of the ITCZ and associated deep penetrative convection. The distinct seasonal migration of the ITCZ across the equator between 8°N during boreal summer and 8°S during austral summer, as derived from precipitation patterns is consistent with similar LCP maxima at similar latitudes.

During boreal winter (DJF) and spring (MAM), the lowest LCP temperatures are concentrated over the maritime continent, where minima approach 182 K. The spatial density of LCPs is also maximized in this region during these seasons, consistent with the findings of Krüger et al. (2008). In contrast, during boreal summer (JJA) and fall (SON), LCP temperatures over the maritime continent are substantially higher, with minima typically exceeding 188 K. The corresponding LCP density is reduced and exhibits a broader spatial distribution across the tropics, particularly over the Indian Ocean, western Pacific, and Atlantic



sectors. Figure 3 reveals a pronounced northward shift of the regions of elevated LCP density during JJA. This displacement aligns with the seasonal migration of the ITCZ and is likely influenced by the onset and progression of the summer monsoon systems, particularly over South and Southeast Asia. The Asian summer monsoon is an important entry pathway into the stratosphere during boreal summer (Wright et al., 2011; Randel et al., 2015; Legras and Bucci, 2020).

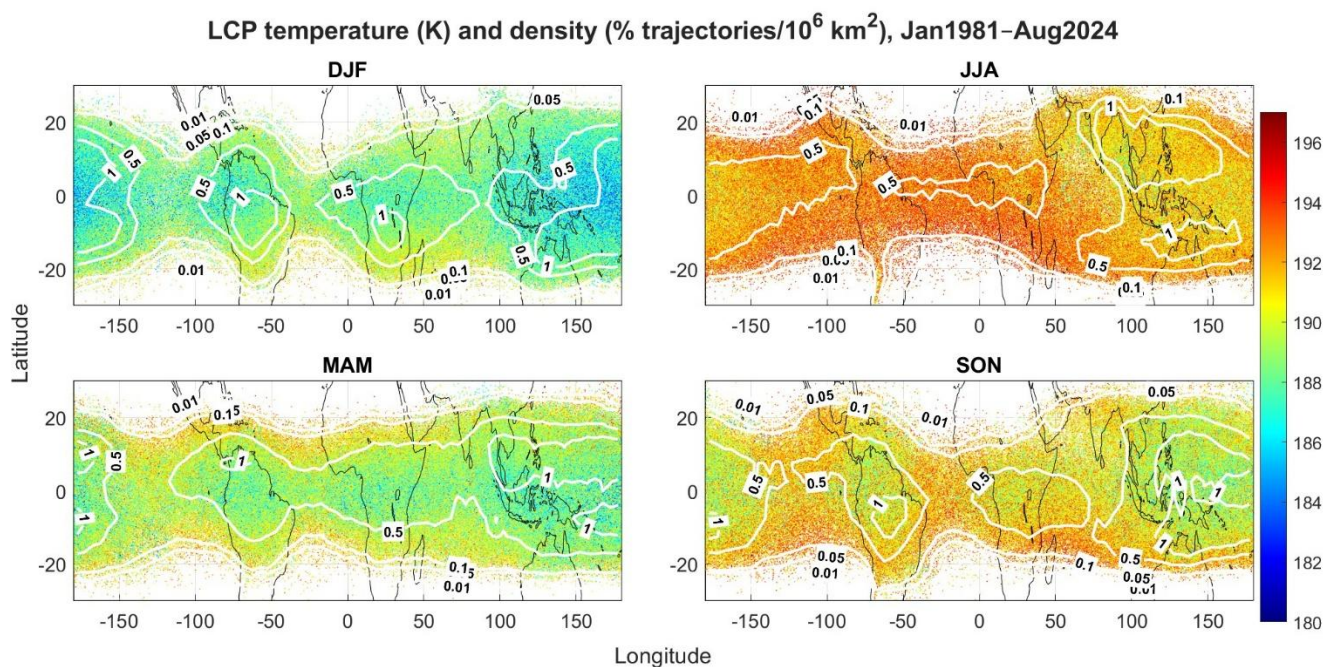
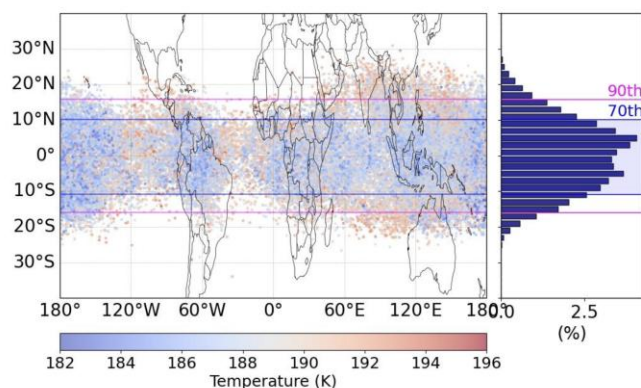


Figure 3. Climatology of the Lagrangian Cold Point (LCP) temperature (K) for different seasons (January 1981 to August 2024). White contour lines display the density of LCP occurrence, expressed as the percentage of trajectories per 1 million km², for DJF, MAM, JJA, and SON.

The interannual variability of LCP temperature and width is illustrated in Fig. 4 based on two contrasting years. In 2016, the LCP distribution is tightly confined to the deep tropics and exhibits substantially lower temperatures, whereas in 2017, LCPs span a much broader latitudinal range and are associated with relatively higher LCP temperatures. The tropical mean ECP temperatures for the two years are 191.16 K in 2016 and 192.13 K in 2017, which is consistent with the observed difference in LCP temperature. The co-occurrence of the lower background temperatures with the narrower LCP distributions suggests that transport pathways in colder years allow for a more efficient sampling of the relatively cold inner tropics when compared to warmer years. This year-to-year contrast is representative of a broader relationship seen throughout the record, as confirmed by the strong correlation between LCP temperature and width ($r = 0.65$; Fig. 6), demonstrating that colder TTL conditions consistently coincide with more confined dehydration pathways.



LCP distribution & latitudinal histogram for 2016



LCP distribution & latitudinal histogram for 2017

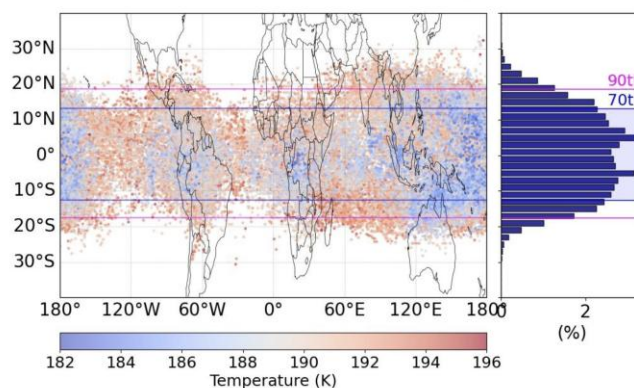


Figure 4. Spatial distribution of Lagrangian Cold Point (LCP) temperature (K) and their latitudinal histograms for the years 2016 (left panel) and 2017 (right panel).

The monthly mean climatological LCP width shows a clear seasonal cycle with values varying between 30° and 37.5° (Fig. 5). The total LCP width exhibits a pronounced seasonal cycle, peaking during boreal summer (June–August) and reaching a minimum during boreal spring (March–May). The dominant contribution to this seasonal variation comes from the NH, which shows a marked expansion of the confinement region in summer months—likely in response to enhanced convection and monsoonal activity that shift TTL dehydration processes poleward. This is also evident from the shift of the high-density LCP region usually found over the Maritime continent and West Pacific towards the northern part of the Maritime continent, the Indian Ocean, and Southeast Asia during boreal summer (Fig. 3). In contrast, the SH width remains steadier throughout the year, with only modest fluctuations that maximise during austral summer (DJF).

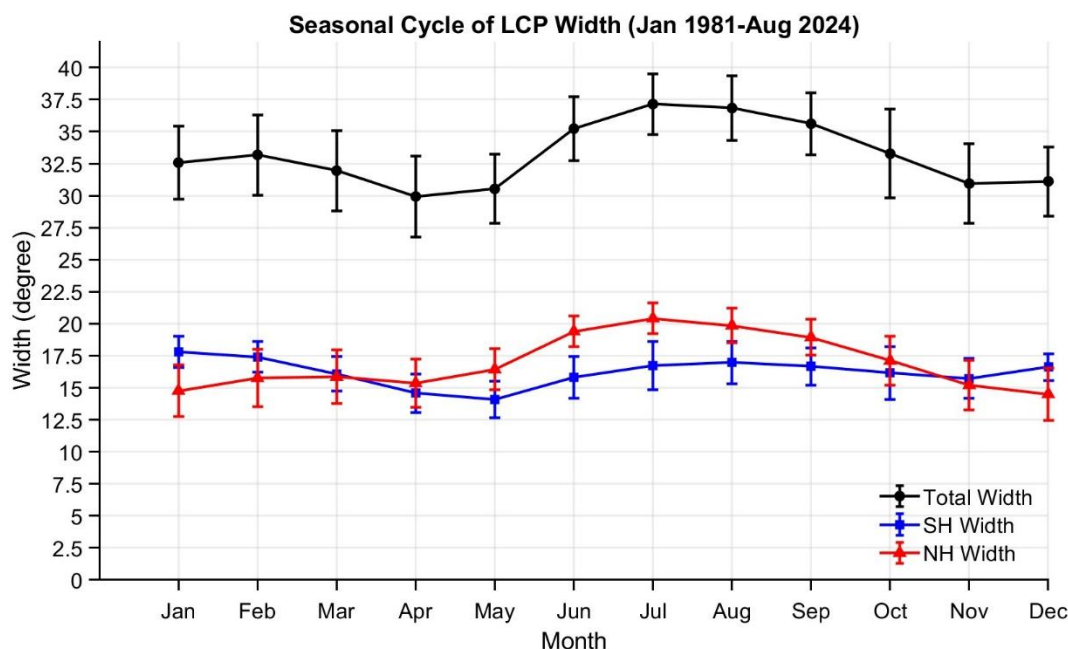


Figure 5. Seasonal cycle of the Lagrangian Cold Point (LCP) region width (°), for January 1981 to August 2024, based on the 90th percentile latitudes. The total LCP width (black line) represents the full meridional extent of the LCP with the Northern Hemisphere (NH) and Southern Hemisphere (SH) edges shown in red and blue, respectively. Vertical error bars indicate ± 1 standard deviation across all years for each month.

5 Long-term trends of LCP temperature and distribution

The long-term linear trends in LCP temperature and its associated meridional width are calculated from the time series of deseasonalized anomalies of the two quantities (Fig. 6). The two time series are highly correlated ($r = 0.65$) with colder years corresponding to narrow LCP distributions and warmer years corresponding to wider LCP distributions. The shared interannual variability shows the impact of the Quasi-Biennial Oscillation (QBO) with positive and negative peaks repeating about every two years.

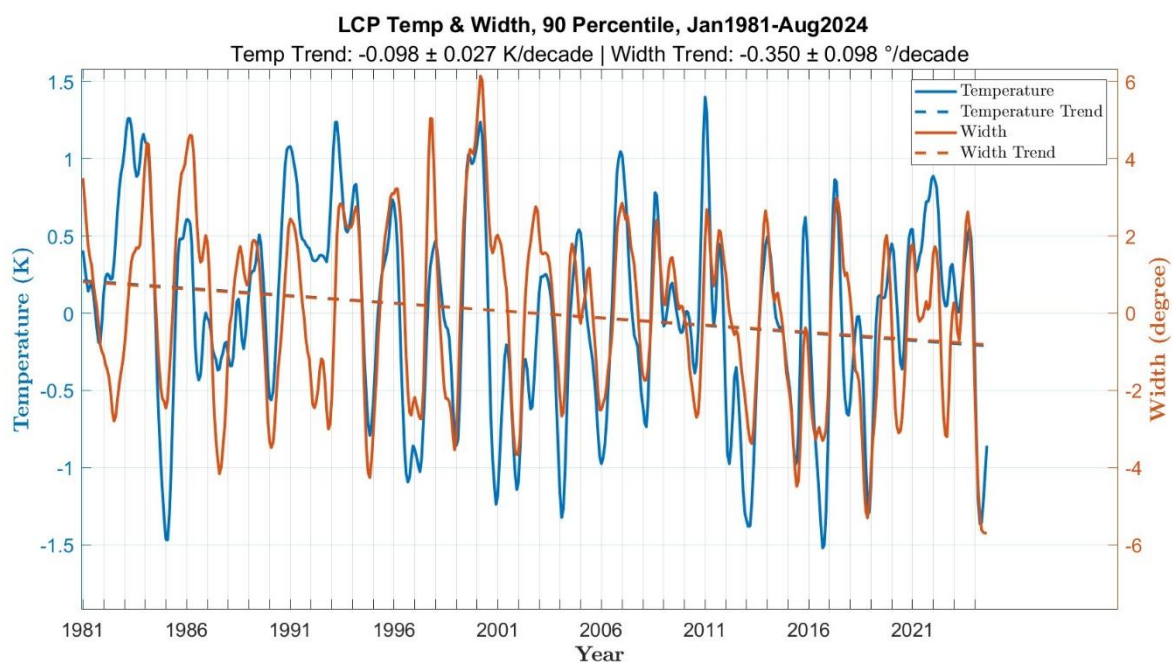


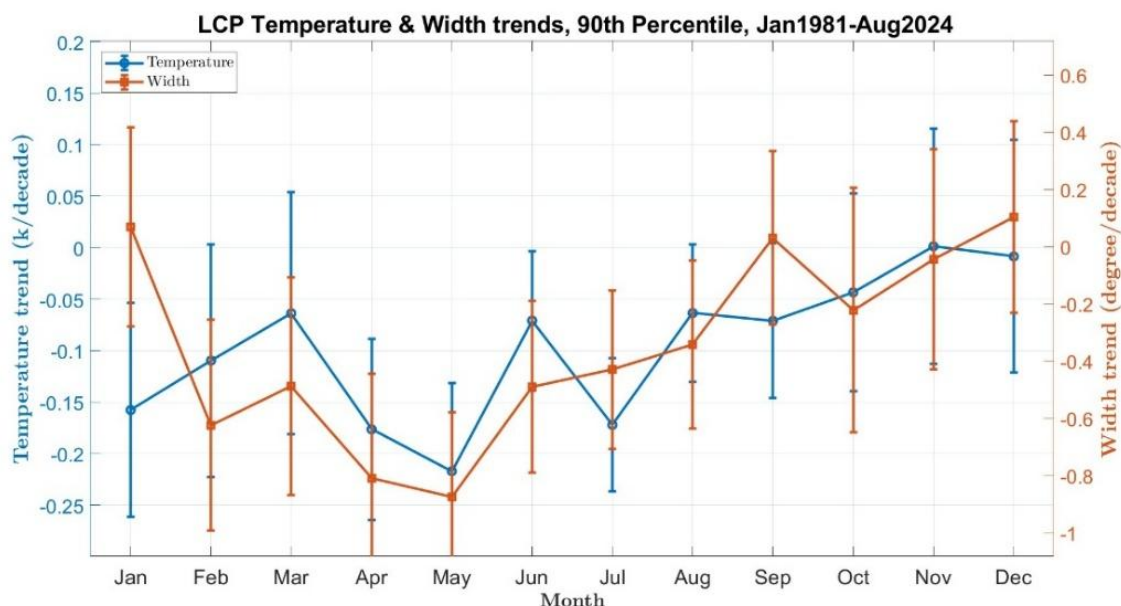
Figure 6. Time series of monthly mean deseasonalized Lagrangian Cold Point (LCP) temperature (K) and width (°) anomalies from January 1981 to August 2024. Linear trends are shown as dashed lines.

In addition to the pronounced interannual variability of the LCP temperature, we find an overall cooling trend of -0.098 ± 0.027 K per decade. This cooling trend suggests a slow change to a more effective dehydration of air masses entering the stratosphere at the cold point. Based on the Clausius-Clapeyron relation, a cooling trend of -0.098 K per decade would result in a trend of -0.065 ± 0.021 ppm/decade. Consistent with the temperature trends, the LCP width shows a significant narrowing trend of $-0.350 \pm 0.098^\circ$ per decade, indicating that the coldest points experienced by ascending air masses in the TTL are becoming increasingly confined to the deep tropics over the January 1981 through August 2024 period. Trends presented here are based on a simple regression without any dynamical or chemical proxies and change only slightly if additional terms such as the QBO or stratospheric aerosol optical depth are included in the regression. LCP pressure shows the same interannual variability as LCP temperature and width and a negative trend of -0.247 ± 0.067 hPa/decade consistent with previous studies that revealed increasing tropopause height (e.g., Seidel et al., 2001; Ladstädter et al., 2025).

Trends in LCP temperature and meridional width display similar seasonal cycles, as shown in Fig. 7. Both variables exhibit predominantly negative trends throughout the year, indicating long-term cooling and narrowing of the LCP. The strongest signals occur during January–May, when cooling approaches -0.20 K per decade and narrowing exceeds -0.6° per decade. From late boreal summer into early winter (August–December), both trends weaken and approach zero, suggesting a seasonal modulation in the long-term evolution of the LCP. Although the trends become less negative during this period, they do not



transition to sustained positive values, indicating that the LCP narrowing and cooling persist year-round, with seasonal variations in magnitude.



335 **Figure 7. Seasonal cycle of the Lagrangian Cold Point (LCP) temperature trends (K per decade) and LCP width trends (° per decade) for the period January 1981 to August 2024. Error bars show the ± 2 sigma confidence intervals of the trend.**

Long-term changes in stratospheric water vapor are strongly impacted by changes in the dehydration temperatures experienced by air parcels ascending through the TTL. In order to analyse whether Lagrangian cold-point calculations are required or if simply Eulerian cold-point time series are sufficient for such analysis, we here compare long-term trends of these two quantities. Figure 8 presents the time series of deseasonalized LCP temperature anomalies, computed from all trajectory-derived cold points within the 90th percentile latitude range, and ECP temperature anomalies (averaged over 15°S–15°N) from January 1981 to August 2024. Both datasets show substantial interannual variability and closely track one another, indicating strong coherence between Lagrangian and Eulerian representations of temperature variability in the TTL. The interannual variability of the LCP time series is larger when compared to ECP, with stronger anomalies during extreme years (e.g., for minimum temperatures in 1985 or maximum temperatures in 2000) consistent with results from Mohr (2013). The larger interannual variability of the LCP temperature time series can result from the occurrence of zonally non-symmetric temperature patterns such as the Matsuno-Gill horseshoe-shaped structure (Nishimoto and Shiotani, 2012) that can enhance Lagrangian mean temperature anomalies more strongly than mean values in the Eulerian framework.

Superimposed linear trends show statistically significant long-term cooling in both metrics: -0.098 ± 0.027 K per decade for the LCP and -0.064 ± 0.020 K per decade for the ECP. The stronger cooling trend in the LCP record indicates that Eulerian temperatures can underestimate cooling at the cold point, and therefore can underestimate changes in stratospheric dehydration, which is controlled by the minimum temperatures encountered along parcel trajectories. This demonstrates that Lagrangian

sampling is necessary to capture the full magnitude of temperature-driven changes in stratospheric water vapor, whereas Eulerian mean temperatures include sectors within the TTL that are irrelevant to dehydration of air entering the stratosphere.

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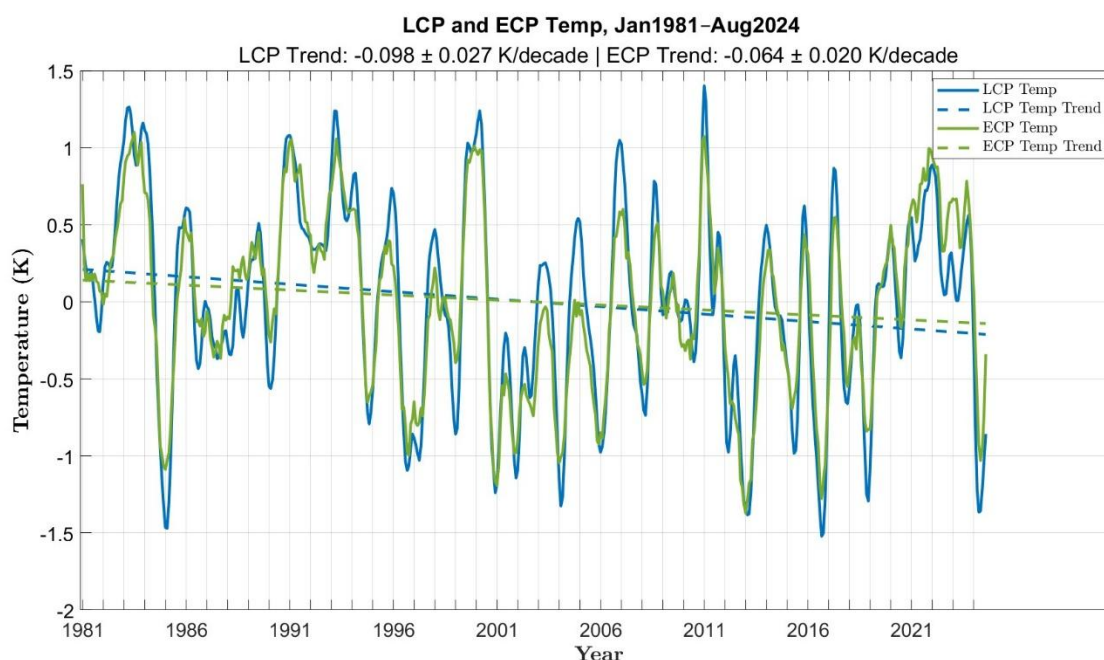


Figure 8. Time series of the deseasonalized Lagrangian Cold Point (LCP) temperature anomalies (K), and Eulerian Cold Point (ECP) temperature anomalies (K) from January 1981 to August 2024. Linear trends are shown as dashed lines.

360 The contribution of background temperature changes to the long-term LCP cooling is isolated from the impact of evolving transport pathways by comparing anomalies of LCP temperature and LCP_varT temperature (Fig. 9). The LCP_varT time series allows us to estimate what the Lagrangian cold-point trend would be if the temperature fields change according to their temporal evolution as given by the reanalysis while the cold-point sampling remains constant from year to year. The two-time series of LCP and LCP_varT temperatures show a very strong match and high correlations. While they share the same signals of interannual variability, their long-term trends differ with LCP cooling at -0.098 ± 0.027 K per decade and LCP_varT cooling at a reduced rate at -0.040 ± 0.025 K per decade. This comparison indicates that if LCP sampling were to remain constant, the LCP cooling trend would be less than half as large, and that temperature changes alone do not account for the full trend in dehydration efficiency at the cold point. This is consistent with the above comparison of LCP and ECP trends that revealed a stronger cooling trend in the LCP record. Both, ECP and LCP_varT trends are solely driven by temperature changes, with the

370 ECP being based on uniformly sampled temperature changes, while the LCP_varT takes the climatological sampling pattern into account.

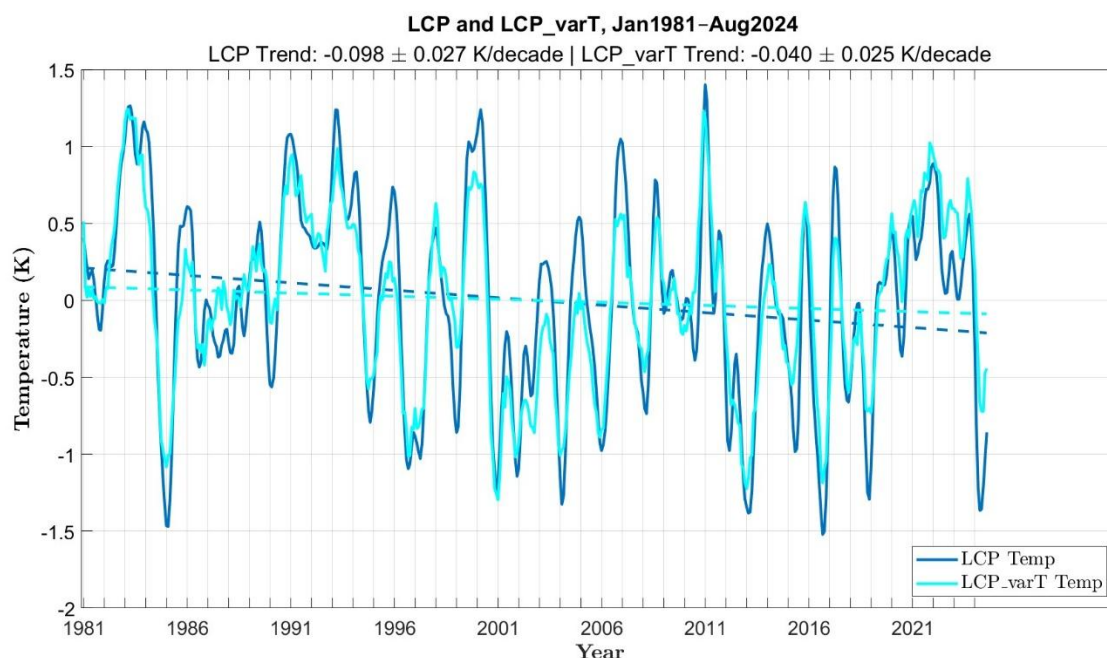


Figure 9. Time series of Lagrangian Cold Point (LCP) temperature anomalies (K, dark blue) and the constant spatial sampling scenario LCP_varT temperature anomalies (K, cyan) from January 1981 to August 2024. Linear trends are shown as dashed lines.

We estimate a general meridional transport distribution to quantify if the general transport pathways throughout the upper TTL, and not just the LCP locations, have narrowed over time. The transport width represents the 90th percentile latitude range of all trajectories situated within a specified pressure layer relative to the climatological monthly mean LCP. Only trajectories contributing to the LCP time series are included in this analysis. Based on the transport width time series, we calculated monthly width anomalies (°) and long-term trends in transport width (° per decade) for January 1981–August 2024. Figure 10 shows the transport width anomalies and trend for one representative layer (LCP to LCP – 5 hPa). This layer exhibits a significant narrowing trend of $-0.177 \pm 0.038^\circ$ per decade, smaller than but consistent with the narrowing of the full LCP width ($-0.350 \pm 0.098^\circ$ per decade). This indicates that air parcels are becoming increasingly confined to a narrower deep-tropical latitude band as they pass the cold point.

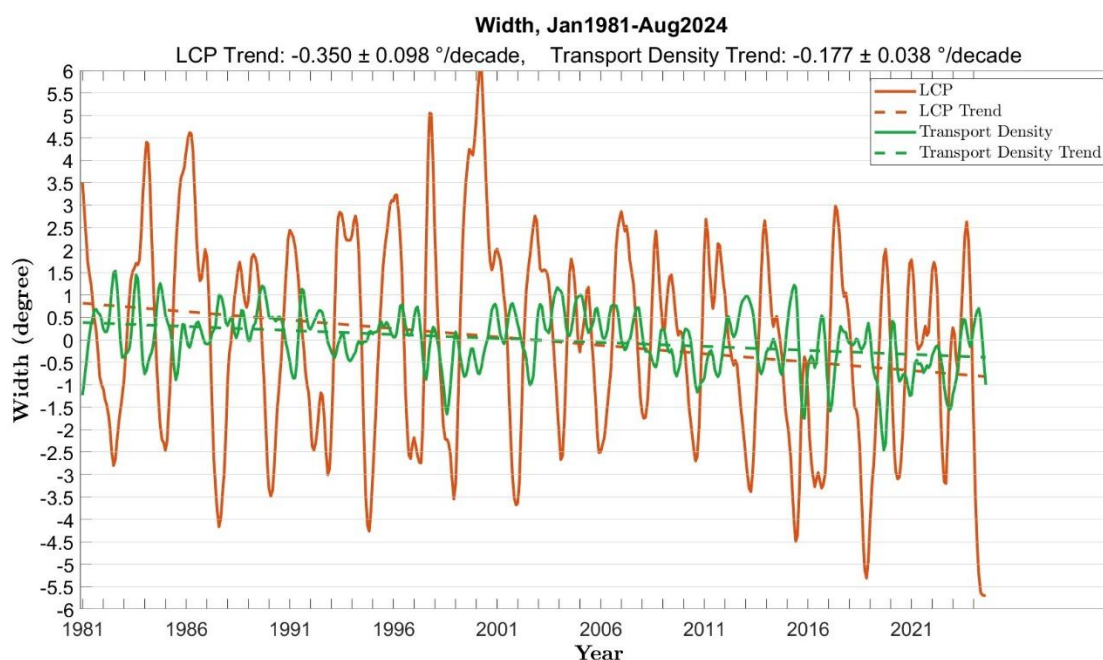


Figure 10. Time series of annual mean deseasonalized of Lagrangian Cold Point (LCP) width anomalies (°, red) and transport width anomalies (°, green) for air parcels located within the layer from the LCP up to 5 hPa above the LCP from January 1981 to August 2024.

390 A full set of transport-width trends for all analysed layers is provided in Table 1. All layers exhibit statistically narrowing, ranging from -0.11 to -0.21 ° per decade, demonstrating that the narrowing signal is not localized to a single pressure layer. The strongest narrowing occurs at the lowest level tested, just below the LCP (LCP + 10 to LCP + 15 hPa, -0.21 ± 0.05 ° per decade). The consistent negative trends of transport width over all layers indicate that transport pathways throughout the upper TTL have become more confined over time. The fact that the LCP narrowing is stronger than the narrowing of the upward

395 transport branch in the individual layers suggests that the transport-temperature patterns determining the LCP positions are affected slightly differently than just the transport patterns themselves.

Table 1. Trends in transport width (° per decade) for different vertical layers relative to the Lagrangian Cold Point (LCP), January 1981–August 2024.

LCP Range (hPa)	Trend in transport width (° per decade)
LCP - 15 hPa to LCP - 10 hPa	-0.13 ± 0.04
LCP - 10 hPa to LCP - 5 hPa	-0.17 ± 0.04
LCP to LCP - 5 hPa	-0.18 ± 0.04



LCP to LCP + 5 hPa	-0.12 ± 0.05
LCP + 5 hPa to LCP + 10 hPa	-0.11 ± 0.05
LCP + 10 to LCP + 15 hPa	-0.21 ± 0.05

400 6 Summary and discussion

This study uses a 44-year Lagrangian trajectory dataset (January 1981–August 2024) to identify where and how air parcels experience the lowest temperatures in the TTL, thereby quantifying the physical pathways that control stratospheric dehydration. A new diagnostic metric—the meridional width of the LCP—is introduced, defined as the latitude range enclosing the 90th percentile of dehydration events. Unlike Eulerian cold-point metrics that sample cold points on a uniform grid, this
 405 Lagrangian framework captures the true transport pathways of air parcels and more accurately represents the processes that regulate water-vapor entry into the stratosphere.

Climatologically, LCP events are strongly concentrated over the Maritime Continent and the western Pacific warm pool, consistent with regions of the coldest TTL temperatures and previous findings (Fueglistaler and Haynes, 2005; Fueglistaler et al., 2005; Krüger et al., 2008). The annual mean width of the LCP region is 33.2°, composed of 16.2° in the SH and 17.0° in
 410 the NH, indicating a nearly symmetric meridional structure across the equator. This ~33° confinement is consistent with earlier trajectory-based analyses, which show that most LCPs and last-dehydration events occur within roughly ±15°–20° of the equator, particularly over regions where deep convection is most intense, and TTL temperatures are coldest (e.g., Fueglistaler et al., 2005; Krüger et al., 2008; Schoeberl et al., 2022). The climatological distribution also reveals a double-peak structure near ±5°, most likely linked to the seasonal migration of the ITCZ and convective activity.

415 The LCP width exhibits a pronounced seasonal cycle. It is largest during boreal summer (June–August), when monsoon systems and associated deep convection extend to higher northern latitudes, and smallest during boreal spring (March–May), when convection and the ITCZ are more tightly confined around the equator. This behavior mirrors the seasonal migration of the ITCZ and the corresponding adjustment of the Hadley circulation, and is consistent with seasonal shifts in tropical rainfall, cold cloud tops, and upper-tropospheric outflow diagnosed from satellite and reanalysis products (e.g., Wodzicki and Rapp,
 420 2016; Byrne et al., 2018; Su et al., 2020). The close agreement between LCP width seasonality and established seasonal patterns in convection and circulation supports the physical interpretability of the trajectory-derived width metric.

LCP temperatures cool at -0.098 ± 0.027 K per decade, a substantially stronger trend than Eulerian cold-point cooling (-0.064 ± 0.020 K per decade) over the same period. This demonstrates that Eulerian diagnostics underestimate the true rate of TTL cooling and associated dehydration. Consistent with our previous Eulerian analysis in Zolghadrshojaee et al. (2024) and
 425 Zolghadrshojaee et al. (2025), which used cold-point temperatures from radiosonde, satellite, and reanalysis data, the strongest



cooling is concentrated in 1981–2001, whereas after 2002 the trend reverses, with a slight warming rather than continued cooling. Fueglistaler et al. (2005) previously showed that Lagrangian cold-point temperatures are systematically colder than Eulerian estimates by ~ 2.5 K and that Eulerian sampling omits key variability relevant to dehydration. Smith et al. (2021) demonstrated that although capturing seasonal and interannual temperature variability is essential for realistically modelling stratospheric water-vapor concentrations, simulation of seasonal variability in transport and of sub-seasonal to sub-monthly temperature variability also plays an important role and cannot be neglected.

A climatological sampling scenario, for which only the temperature changes from year to year, shows that temperature changes alone yield negative trends, but only about half the magnitude of the full LCP trend. This confirms that both changes in temperature and changes in sampling location contribute to long-term dehydration variability. Smith et al. (2021) quantified the relative roles of variations in TTL temperatures and transport in determining seasonal and interannual variations of stratospheric water vapor. Using trajectory calculations, they showed that variations in temperatures, rather than transport, dominate interannual variability, typically explaining more than 70% of the variance—including extreme events such as the 2000 stratospheric water-vapor drop. Similarly, seasonal variations in temperatures, rather than transport, are the dominant driver of the annual cycle in lower-stratospheric water-vapor concentrations in both models and reanalysis; however, seasonal variations in transport play a non-negligible role by reducing the seasonal-cycle maximum and lowering the annual range by approximately 30%. Together, these findings highlight that both temperature variability and transport variability must be accounted for when interpreting variability and trends of the dehydration processes.

The LCP exhibits a statistically significant narrowing trend of $-0.350 \pm 0.098^\circ$ per decade from January 1981 to August 2024. Vertical layers ± 15 hPa around the LCP that contain all trajectories transported into the stratosphere also narrow (-0.11 to -0.21° per decade), demonstrating a contraction of the upward motion pathways throughout the upper TTL. The greater narrowing of the LCPs compared with the upward transport branch in individual layers suggests that the combined transport–temperature patterns governing LCP positions respond somewhat differently from the transport patterns alone.

The dynamical changes of the LCP are consistent with recently reported narrowing of the ITCZ. Observational and modelling studies show that the deep-convective ITCZ band has contracted in recent decades (Wodzicki and Rapp, 2016; Byrne et al., 2018; Su et al., 2020), even as the broader tropical belt—defined by Hadley cell edges, subtropical jets, and tropopause height—has expanded (e.g., Staten et al., 2018; Staten et al., 2020). A 1979–2017 time series of ITCZ width for the Atlantic and Pacific basins constructed from GPCP precipitation data shows considerable narrowing of about 22 to 38 km/decade (Byrne et al., 2018). The ITCZ narrowing corresponds to an increase in precipitation intensity which can be explained by a shift towards more frequent organized deep convection near the ITCZ centre and less frequent convection along its edges (Tan et al., 2015).

The LCP width trends of $-0.350 \pm 0.098^\circ$ per decade are very similar to the observed ITCZ trends of 22 to 38 km/decade found for a similar time period. While a targeted analysis demonstrating the connection between the LCP and ITCZ locations is beyond the scope of this study, our results provide some indication that the LCP-based dehydration pathways narrow in a manner consistent with the tightening of the ITCZ and associated deep convection. The observed narrowing of LCP width,



460 therefore, possibly provides a vertical extension of ITCZ narrowing into the stratospheric entry region, suggesting that dehydration at stratospheric entry could be impacted by tropospheric dynamical trends. At the same time the LCP width is impacted by stratospheric processes as apparent from the clear QBO signal in the LCP width.

Collectively, these results show that pathways for stratospheric dehydration are becoming geographically more focused rather than broader. The LCP is cooling and narrowing, increasing the selectivity of entry pathways into the stratosphere and implying
465 potential impacts on stratospheric water vapor and radiative forcing. The demonstrated differences between Lagrangian and Eulerian diagnostics emphasize that trajectory-based tools are essential for accurately assessing long-term variability, climate sensitivity, and future changes in TTL dehydration processes. More broadly, the combination of tropospheric tropical-belt widening, ITCZ narrowing, and LCP-width narrowing points to a vertically structured adjustment of tropical circulation in a warming climate, in which the outer margins of the tropics expand while the deep-convective and dehydration core becomes
470 more tightly confined.

Data availability

MERRA-2 reanalysis data are publicly available from the NASA Goddard Earth Sciences Data and Information Services Center (GMAO, 2015a, b, c, d). The trajectory dataset and the derived Lagrangian cold-point data used in this study are available from the authors upon request.

475 Author contributions

MZ performed the analysis and wrote the manuscript. ST developed the research question and supervised the study. SMD provided the trajectory dataset and contributed to the interpretation of the results. KK contributed to the interpretation of the results. All authors discussed the results and commented on the manuscript.

Competing interests

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

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References

- Avery, M. A., Davis, S. M., Rosenlof, K. H., Ye, H., and Dessler, A. E.: Large anomalies in lower stratospheric water vapour and ice during the 2015–2016 El Niño, *Nat. Geosci.*, 10, 405–409, <https://doi.org/10.1038/ngeo2961>, 2017.
- Birner, T.: Recent widening of the tropical belt from global tropopause statistics: Sensitivities, *J. Geophys. Res.*, 115, D23 109, <https://doi.org/10.1029/2010JD014664>, 2010.
- 490 Bonazzola, M. and Haynes, P. H.: A trajectory-based study of the tropical tropopause region, *J. Geophys. Res.*, 109, 2003JD004356, <https://doi.org/10.1029/2003JD004356>, 2004.
- Brewer, A. W.: Evidence for a world circulation provided by the measurements of helium and water vapour distribution in the stratosphere, *Q. J. Roy. Meteor. Soc.*, 75, 351–363, <https://doi.org/10.1002/qj.49707532603>, 1949.
- 495 Byrne, M. P. and Schneider, T.: Energetic constraints on the width of the intertropical convergence zone, *J. Climate*, 29, 4709–4721, <https://doi.org/10.1175/JCLI-D-15-0767.1>, 2016.
- Byrne, M. P., Pendergrass, A. G., Rapp, A. D., and Wodzicki, K. R.: Response of the intertropical convergence zone to climate change: Location, width, and strength, *Curr. Clim. Change Rep.*, 4, 355–370, <https://doi.org/10.1007/s40641-018-0110-5>, 2018.
- 500 Corti, T., Luo, B. P., de Reus, M., Brunner, D., Cairo, F., Mahoney, M. J., Martucci, G., Matthey, R., Mitev, V., dos Santos, F. H., Schiller, C., Shur, G., Sitnikov, N. M., Spelten, N., Vössing, H. J., Borrmann, S., and Peter, T.: Unprecedented evidence for deep convection hydrating the tropical stratosphere, *Geophys. Res. Lett.*, 35, 2008GL033641, <https://doi.org/10.1029/2008GL033641>, 2008.
- Davis, N. A. and Davis, S. M.: Reconciling Hadley cell expansion trend estimates in reanalyses, *Geophys. Res. Lett.*, 45, 11439–11446, <https://doi.org/10.1029/2018GL079593>, 2018.
- 505 Davis, S. M. and Rosenlof, K. H.: A multidiagnostic intercomparison of tropical-width time series using reanalyses and satellite observations, *J. Climate*, 25, 1061–1078, <https://doi.org/10.1175/JCLI-D-11-00127.1>, 2012.
- Donohoe, A., Atwood, A. R., and Byrne, M. P.: Controls on the width of tropical precipitation and its contraction under global warming, *Geophys. Res. Lett.*, 46, 9958–9967, <https://doi.org/10.1029/2019GL082969>, 2019.
- 510 Folkins, I., Loewenstein, M., Podolske, J., Oltmans, S. J., and Proffitt, M.: A barrier to vertical mixing at 14 km in the tropics: Evidence from ozonesondes and aircraft measurements, *J. Geophys. Res.*, 104, 22095–22102, <https://doi.org/10.1029/1999JD900404>, 1999.
- Frierson, D. M. W., Lu, J., and Chen, G.: Width of the Hadley cell in simple and comprehensive general circulation models, *Geophys. Res. Lett.*, 34, 2007GL031115, <https://doi.org/10.1029/2007GL031115>, 2007.



- 515 Fueglistaler, S. and Haynes, P. H.: Control of interannual and longer-term variability of stratospheric water vapor, *J. Geophys. Res.*, 110, 2005JD006019, <https://doi.org/10.1029/2005JD006019>, 2005.
 Fueglistaler, S., Bonazzola, M., Haynes, P. H., and Peter, T.: Stratospheric water vapor predicted from the Lagrangian temperature history of air entering the stratosphere in the tropics, *J. Geophys. Res.*, 110, 2004JD005516, <https://doi.org/10.1029/2004JD005516>, 2005.
- 520 Fueglistaler, S., Dessler, A. E., Dunkerton, T. J., Folkins, I., Fu, Q., and Mote, P. W.: Tropical tropopause layer, *Rev. Geophys.*, 47, 2008RG000267, <https://doi.org/10.1029/2008RG000267>, 2009.
 Gelaro, R., McCarty, W., Suárez, M. J., Todling, R., Molod, A., Takacs, L., Randles, C. A., Darmenov, A., Bosilovich, M. G., Reichle, R., Wargan, K., Coy, L., Cullather, R., Draper, C., Akella, S., Buchard, V., Conaty, A., Da Silva, A. M., Gu, W., Kim, G.-K., Koster, R., Lucchesi, R., Merkova, D., Nielsen, J. E., Partyka, G., Pawson, S., Putman, W., Rienecker, M., Schubert, S.
 525 D., Sienkiewicz, M., and Zhao, B.: The Modern-Era Retrospective Analysis for Research and Applications, Version 2 (MERRA-2), *J. Climate*, 30, 5419–5454, <https://doi.org/10.1175/JCLI-D-16-0758.1>, 2017.
 Global Modeling and Assimilation Office (GMAO): MERRA-2 inst6_3d_ana_Nv: 3d, 6-Hourly, Instantaneous, Model-Level, Analysis, Analyzed Meteorological Fields V5.12.4, Goddard Earth Sciences Data and Information Services Center (GES DISC), Greenbelt, MD, USA [data set], <https://doi.org/10.5067/IUUF4WB9FT4W>, 2015a.
- 530 Global Modeling and Assimilation Office (GMAO): MERRA-2 inst3_3d_asm_Nv: 3d, 3-Hourly, Instantaneous, Model-Level, Assimilation, Assimilated Meteorological Fields V5.12.4, Goddard Earth Sciences Data and Information Services Center (GES DISC), Greenbelt, MD, USA [data set], <https://doi.org/10.5067/WWQSQXQ8IVFW8>, 2015b.
 Global Modeling and Assimilation Office (GMAO): MERRA-2 tavg3_3d_tdt_Np: 3d, 3-Hourly, Time-Averaged, Pressure-Level, Assimilation, Temperature Tendencies V5.12.4, Goddard Earth Sciences Data and Information Services Center (GES
 535 DISC), Greenbelt, MD, USA [data set], <https://doi.org/10.5067/9NCR9DDDOPFI>, 2015c.
 Global Modeling and Assimilation Office (GMAO): MERRA-2 inst3_3d_asm_Np: 3d, 3-Hourly, Instantaneous, Pressure-Level, Assimilation, Assimilated Meteorological Fields V5.12.4, Goddard Earth Sciences Data and Information Services Center (GES DISC), Greenbelt, MD, USA [data set], <https://doi.org/10.5067/QBZ6MG944HW0>, 2015d.
 Grise, K. M., Davis, S. M., Simpson, I. R., Waugh, D. W., Fu, Q., Allen, R. J., Rosenlof, K. H., Ummenhofer, C. C.,
 540 Karnauskas, K. B., Maycock, A. C., Quan, X.-W., Birner, T., and Staten, P. W.: Recent tropical expansion: Natural variability or forced response?, *J. Climate*, 32, 1551–1571, <https://doi.org/10.1175/JCLI-D-18-0444.1>, 2019.
 Harzer, F., Garny, H., Davis, S., and Birner, T.: Probing the suitability of meridional stratospheric ozone gradients for inferring interannual variability and trends of the subtropical jet stream, *J. Climate*, 38, 2571–2588, <https://doi.org/10.1175/JCLI-D-24-0530.1>, 2025.
- 545 Hegglin, M. I., Plummer, D. A., Shepherd, T. G., Scinocca, J. F., Anderson, J., Froidevaux, L., Funke, B., Hurst, D., Rozanov, A., Urban, J., von Clarmann, T., Walker, K. A., Wang, H. J., Tegtmeier, S., and Weigel, K.: Vertical structure of stratospheric water vapour trends derived from merged satellite data, *Nat. Geosci.*, 7, 768–776, <https://doi.org/10.1038/ngeo2236>, 2014.



- Highwood, E. J. and Hoskins, B. J.: The tropical tropopause, *Q. J. Roy. Meteor. Soc.*, 124, 1579–1604, <https://doi.org/10.1002/qj.49712454911>, 1998.
- 550 Holton, J. R. and Gettelman, A.: Horizontal transport and the dehydration of the stratosphere, *Geophys. Res. Lett.*, 28, 2799–2802, <https://doi.org/10.1029/2001GL013148>, 2001.
- Holton, J. R., Haynes, P. H., McIntyre, M. E., Douglass, A. R., Rood, R. B., and Pfister, L.: Stratosphere–troposphere exchange, *Rev. Geophys.*, 33, 403–439, <https://doi.org/10.1029/95RG02097>, 1995.
- Hu, Y. and Fu, Q.: Observed poleward expansion of the Hadley circulation since 1979, *Atmos. Chem. Phys.*, 7, 5229–5236, <https://doi.org/10.5194/acp-7-5229-2007>, 2007.
- 555 Jain, A. R., Panwar, V., Johny, C. J., Mandal, T. K., Rao, V. R., Gautam, R., and Dhaka, S. K.: Occurrence of extremely low cold point tropopause temperature during summer monsoon season: ARMEX campaign and CHAMP and COSMIC satellite observations, *J. Geophys. Res.*, 116, D03102, <https://doi.org/10.1029/2010JD014340>, 2011.
- Johanson, C. M. and Fu, Q.: Hadley cell widening: Model simulations versus observations, *J. Climate*, 22, 2713–2725, <https://doi.org/10.1175/2008JCLI2620.1>, 2009.
- 560 Kim, J. and Son, S.-W.: Tropical cold-point tropopause: Climatology, seasonal cycle, and intraseasonal variability derived from COSMIC GPS radio occultation measurements, *J. Climate*, 25, 5343–5360, <https://doi.org/10.1175/JCLI-D-11-00554.1>, 2012.
- Krüger, K., Tegtmeier, S., and Rex, M.: Long-term climatology of air mass transport through the Tropical Tropopause Layer (TTL) during NH winter, *Atmos. Chem. Phys.*, 8, 813–823, <https://doi.org/10.5194/acp-8-813-2008>, 2008.
- Krüger, K., Tegtmeier, S., and Rex, M.: Variability of residence time in the Tropical Tropopause Layer during Northern Hemisphere winter, *Atmos. Chem. Phys.*, 9, 6717–6725, <https://doi.org/10.5194/acp-9-6717-2009>, 2009.
- Ladstädter, F., Stocker, M., Scher, S., and Steiner, A. K.: Observed changes in the temperature and height of the globally resolved lapse rate tropopause, *Atmos. Chem. Phys.*, 25, 16053–16062, <https://doi.org/10.5194/acp-25-16053-2025>, 2025.
- 570 Lau, W. K. M. and Kim, K.-M.: Robust Hadley circulation changes and increasing global dryness due to CO₂ warming from CMIP5 model projections, *P. Natl. Acad. Sci. USA*, 112, 3630–3635, <https://doi.org/10.1073/pnas.1418682112>, 2015.
- Legras, B. and Bucci, S.: Confinement of air in the Asian monsoon anticyclone and pathways of convective air to the stratosphere during the summer season, *Atmos. Chem. Phys.*, 20, 11045–11064, <https://doi.org/10.5194/acp-20-11045-2020>, 2020.
- Lu, J., Vecchi, G. A., and Reichler, T.: Expansion of the Hadley cell under global warming, *Geophys. Res. Lett.*, 34, 2006GL028443, <https://doi.org/10.1029/2006GL028443>, 2007.
- 575 Manney, G. L. and Hegglin, M. I.: Seasonal and regional variations of long-term changes in upper-tropospheric jets from reanalyses, *J. Climate*, 31, 423–448, <https://doi.org/10.1175/JCLI-D-17-0303.1>, 2018.
- Marcy, T. P., Popp, P. J., Gao, R. S., Fahey, D. W., Ray, E. A., Richard, E. C., Thompson, T. L., Atlas, E. L., Loewenstein, M., Wofsy, S. C., Park, S., Weinstock, E. M., Swartz, W. H., and Mahoney, M. J.: Measurements of trace gases in the tropical tropopause layer, *Atmos. Environ.*, 41, 7253–7261, <https://doi.org/10.1016/j.atmosenv.2007.05.032>, 2007.
- 580



- Mohr, V.: Changes of the tropical tropopause layer in reanalysis and climate model data, PhD thesis, Christian-Albrechts-Universität zu Kiel, Kiel, Germany, 159 pp., 2013.
- Nishimoto, E. and Shiotani, M.: Seasonal and interannual variability in the temperature structure around the tropical tropopause and its relationship with convective activities, *J. Geophys. Res.*, 117, D02104, <https://doi.org/10.1029/2011JD016936>, 2012.
- 585 Proud, S. R. and Bachmeier, S.: Record-low cloud temperatures associated with a tropical deep convective event, *Geophys. Res. Lett.*, 48, e2020GL092261, <https://doi.org/10.1029/2020GL092261>, 2021.
- Randel, W. and Park, M.: Diagnosing observed stratospheric water vapor relationships to the cold point tropical tropopause, *J. Geophys. Res.-Atmos.*, 124, 7018–7033, <https://doi.org/10.1029/2019JD030648>, 2019.
- Randel, W. J., Wu, F., and Rivera Ríos, W.: Thermal variability of the tropical tropopause region derived from GPS/MET
 590 observations, *J. Geophys. Res.*, 108, 4024, <https://doi.org/10.1029/2002JD002595>, 2003.
- Randel, W. J., Zhang, K., and Fu, R.: What controls stratospheric water vapor in the NH summer monsoon regions?, *J. Geophys. Res.-Atmos.*, 120, 7988–8001, <https://doi.org/10.1002/2015JD023622>, 2015.
- Santer, B. D., Wigley, T. M. L., Boyle, J. S., Gaffen, D. J., Hnilo, J. J., Nychka, D., Parker, D. E., and Taylor, K. E.: Statistical significance of trends and trend differences in layer-average atmospheric temperature time series, *J. Geophys. Res.*, 105, 7337–
 595 7356, <https://doi.org/10.1029/1999JD901105>, 2000.
- Schiro, K. A., Su, H., Wang, Y., Langenbrunner, B., Jiang, J. H., and Neelin, J. D.: Relationships between tropical ascent and high cloud fraction changes with warming revealed by perturbation physics experiments in CAM5, *Geophys. Res. Lett.*, 46, 10112–10121, <https://doi.org/10.1029/2019GL083026>, 2019.
- Schiller, C., Grooß, J. U., Konopka, P., Plöger, F., Silva dos Santos, F. H., and Spelten, N.: Hydration and dehydration at the
 600 tropical tropopause, *Atmos. Chem. Phys.*, 9, 9647–9660, <https://doi.org/10.5194/acp-9-9647-2009>, 2009.
- Schoeberl, M. R. and Dessler, A. E.: Dehydration of the stratosphere, *Atmos. Chem. Phys.*, 11, 8433–8446, <https://doi.org/10.5194/acp-11-8433-2011>, 2011.
- Schoeberl, M. R., Dessler, A. E., and Wang, T.: Simulation of stratospheric water vapor and trends using three reanalyses, *Atmos. Chem. Phys.*, 12, 6475–6487, <https://doi.org/10.5194/acp-12-6475-2012>, 2012.
- 605 Schoeberl, M. R., Jensen, E. J., Pfister, L., Ueyama, R., Avery, M., and Dessler, A. E.: Convective hydration of the upper troposphere and lower stratosphere, *J. Geophys. Res.-Atmos.*, 123, 4583–4593, <https://doi.org/10.1029/2018JD028286>, 2018.
- Schoeberl, M. R., Ueyama, R., and Pfister, L.: A Lagrangian view of seasonal stratosphere–troposphere exchange, *J. Geophys. Res.-Atmos.*, 127, e2022JD036772, <https://doi.org/10.1029/2022JD036772>, 2022.
- Seidel, D. J. and Randel, W. J.: Recent widening of the tropical belt: Evidence from tropopause observations, *J. Geophys. Res.*,
 610 112, 2007JD008861, <https://doi.org/10.1029/2007JD008861>, 2007.
- Seidel, D. J., Ross, R. J., Angell, J. K., and Reid, G. C.: Climatological characteristics of the tropical tropopause as revealed by radiosondes, *J. Geophys. Res.*, 106, 7857–7878, <https://doi.org/10.1029/2000JD900837>, 2001.



- Smith, J. W., Haynes, P. H., Maycock, A. C., Butchart, N., and Bushell, A. C.: Sensitivity of stratospheric water vapour to variability in tropical tropopause temperatures and large-scale transport, *Atmos. Chem. Phys.*, 21, 2469–2489, 615 <https://doi.org/10.5194/acp-21-2469-2021>, 2021.
- SPARC: SPARC Reanalysis Intercomparison Project (S-RIP) Final Report, edited by: Fujiwara, M., Manney, G. L., Gray, L. J., and Wright, J. S., SPARC Report No. 10, WCRP-6/2021, <https://doi.org/10.17874/800dee57d13>, 2022.
- Staten, P. W., Lu, J., Grise, K. M., Davis, S. M., and Birner, T.: Re-examining tropical expansion, *Nat. Clim. Change*, 8, 768–775, <https://doi.org/10.1038/s41558-018-0246-2>, 2018.
- 620 Staten, P. W., Grise, K. M., Davis, S. M., Karnauskas, K. B., Waugh, D. W., Maycock, A. C., Fu, Q., Cook, K., Adam, O., Simpson, I. R., Allen, R. J., Rosenlof, K., Chen, G., Ummenhofer, C. C., Quan, X.-W., Kossin, J. P., Davis, N. A., and Son, S.-W.: Tropical widening: From global variations to regional impacts, *B. Am. Meteorol. Soc.*, 101, E897–E904, <https://doi.org/10.1175/BAMS-D-19-0047.1>, 2020.
- Steinwagner, J., Fueglistaler, S., Stiller, G., von Clarmann, T., Kiefer, M., Borsboom, P.-P., van Delden, A., and Röckmann, 625 T.: Tropical dehydration processes constrained by the seasonality of stratospheric deuterated water, *Nat. Geosci.*, 3, 262–266, <https://doi.org/10.1038/ngeo822>, 2010.
- Su, H., Jiang, J. H., Neelin, J. D., Shen, T. J., Zhai, C., Yue, Q., Wang, Z., Huang, L., Choi, Y.-S., Stephens, G. L., and Yung, Y. L.: Tightening of tropical ascent and high clouds key to precipitation change in a warmer climate, *Nat. Commun.*, 8, 15771, <https://doi.org/10.1038/ncomms15771>, 2017.
- 630 Su, H., Wu, L., Zhai, C., Jiang, J. H., Neelin, J. D., and Yung, Y. L.: Observed tightening of tropical ascent in recent decades and linkage to regional precipitation changes, *Geophys. Res. Lett.*, 47, e2019GL085809, <https://doi.org/10.1029/2019GL085809>, 2020.
- Tan, J., Jakob, C., Rossow, W. B., and Tselioudis, G.: Increases in tropical rainfall driven by changes in frequency of organized deep convection, *Nature*, 519, 451–454, <https://doi.org/10.1038/nature14339>, 2015.
- 635 Tegtmeier, S., Anstey, J., Davis, S., Dragani, R., Harada, Y., Ivanciu, I., Pilch Kedzierski, R., Krüger, K., Legras, B., Long, C., Wang, J. S., Wargan, K., and Wright, J. S.: Temperature and tropopause characteristics from reanalyses data in the tropical tropopause layer, *Atmos. Chem. Phys.*, 20, 753–770, <https://doi.org/10.5194/acp-20-753-2020>, 2020.
- Tuck, A. F., Baumgardner, D., Chan, K. R., Dye, J. E., Elkins, J. W., Hovde, S. J., Kelly, K. K., Loewenstein, M., Margitan, J. J., May, R. D., Podolske, J. R., Proffitt, M. H., Rosenlof, K. H., Smith, W. L., Webster, C. R., and Wilson, J. C.: The Brewer– 640 Dobson circulation in the light of high altitude in situ aircraft observations, *Q. J. Roy. Meteor. Soc.*, 123, 1–69, <https://doi.org/10.1256/smsqj.53701>, 1997.
- Wang, T., Dessler, A. E., Schoeberl, M. R., Randel, W. J., and Kim, J.-E.: The impact of temperature vertical structure on trajectory modeling of stratospheric water vapor, *Atmos. Chem. Phys.*, 15, 3517–3526, <https://doi.org/10.5194/acp-15-3517-2015>, 2015.



- 645 Wang, X., Dessler, A. E., Schoeberl, M. R., Yu, W., and Wang, T.: Impact of convectively lofted ice on the seasonal cycle of water vapor in the tropical tropopause layer, *Atmos. Chem. Phys.*, 19, 14621–14636, <https://doi.org/10.5194/acp-19-14621-2019>, 2019.
- Wodzicki, K. R. and Rapp, A. D.: Long-term characterization of the Pacific ITCZ using TRMM, GPCP, and ERA-Interim, *J. Geophys. Res.-Atmos.*, 121, 3153–3170, <https://doi.org/10.1002/2015JD024458>, 2016.
- 650 Wright, J. S., Fu, R., Fueglistaler, S., Liu, Y. S., and Zhang, Y.: The influence of summertime convection over Southeast Asia on water vapor in the tropical stratosphere, *J. Geophys. Res.-Atmos.*, 116, D12302, <https://doi.org/10.1029/2010JD015416>, 2011.
- Zhou, X. and Holton, J. R.: Intraseasonal variations of tropical cold-point tropopause temperatures, *J. Climate*, 15, 1460–1473, [https://doi.org/10.1175/1520-0442\(2002\)015<1460:IVOTCP>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<1460:IVOTCP>2.0.CO;2), 2002.
- 655 Zolghadrshojaei, M., Tegtmeier, S., Davis, S. M., and Pilch Kedzierski, R.: Variability and long-term changes in tropical cold-point temperature and water vapor, *Atmos. Chem. Phys.*, 24, 7405–7419, <https://doi.org/10.5194/acp-24-7405-2024>, 2024.
- Zolghadrshojaei, M., Tegtmeier, S., Davis, S. M., Pilch Kedzierski, R., and Haimberger, L.: Shift in cold-point tropopause trends derived from radiosonde, satellite and reanalysis data, *Atmos. Chem. Phys.*, 25, 12213–12232, <https://doi.org/10.5194/acp-25-12213-2025>, 2025.