Moisture transport by convective overshoots in the tropical tropopause layer

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Introduction

The tropical tropopause layer (TTL) is a transitional layer in the tropics between the troposphere, where temperature decreases with height, and the stratosphere, where temperature increases with height. The TTL is important as a gateway to the wider stratosphere; the Brewer–Dobson circulation (BDC) carries air from the upper part of the TTL into the tropical and then extratropical stratosphere on timescales of a few years (Butchart, 2014), significantly influencing the global stratospheric composition. The TTL is also interesting as a transition region between two distinct dynamical regimes; it can be identified as the region where the significant longitudinal variations in tropospheric dynamics manifest themselves in the more zonally uniform equatorial stratosphere (Fueglistaler *et al*., 2009).

Definitions of the TTL structure vary. We adopt the Fueglistaler *et al*. (2009) definition as shown in Figure 1, with the lower boundary of the TTL at 14.5km, above the height of the major convective outflow, and the upper boundary at 18.5km. There is a temperature minimum at the cold-point tropopause (CPT) around 16.5km. We refer to the region above the CPT as the upper TTL. Within the TTL, above the level of zero radiative heating (LZRH) around 15km, there is net radiative heating and corresponding large-scale slow upwelling which forms the start of the BDC. Below the LZRH, in the air outside of convective clouds, there is radiative cooling and correspondingly large-scale subsidence. Outside of convective regions, motion in the upper TTL is quasi-horizontal, with air parcels travelling significant horizontal distances during their slow upwelling over several weeks.

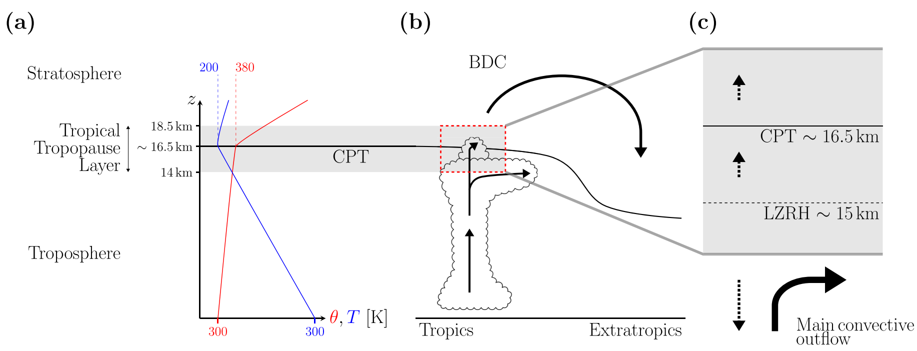

Figure 1. Schematic diagram of convective overshooting and the tropical tropopause layer (TTL). (a) Layers of the atmospheric structure and approximate upper and lower bounds of the TTL and cold-point tropopause (CPT) height, alongside an illustrative vertical profile of potential temperature θ and temperature T with approximate values (in K) at the surface and CPT. (b) The solid line indicates latitudinal variation of the CPT height, grey shading the TTL, and arrows the motion in deep convection. The main convective outflow is in the upper troposphere, while convective overshoots penetrate into the TTL and occasionally cross the CPT (red dashed box, indicating the region considered in Figure 5). The large arrow illustrates transport of air from the TTL into the extratropical stratosphere by the Brewer-Dobson circulation (BDC). (c) This shows TTL structure and vertical flow outside convection, slow upwelling (radiative heating) above the level of zero radiative heating (LZRH) and subsidence (radiative cooling) below the LZRH and the main convective outflow below the TTL (solid arrow). The CPT lies above the LZRH and constrains the water vapour concentration of air parcels that slowly rise through the TTL.

Note that we refer to the region above the TTL as the tropical lower stratosphere (TLS): many studies refer to the region above the CPT, including the upper TTL, as part of the lower stratosphere.

TTL processes play an important role in determining water vapour concentrations in the stratosphere. The concentration of water vapour in a parcel of air is limited by the saturation vapour concentration which, according to the Clausius–Clapeyron relation, decreases rapidly with temperature. The very low temperatures at the tropical CPT reduce water vapour concentrations to the order of a few parts per million by volume (ppmv). The fact that the CPT is above the LZRH is an important aspect of vertical water vapour transport and dehydration in the TTL; the majority of the air that penetrates above the LZRH is likely to ascend into the stratosphere and the same is true of air passing through the CPT, meaning that temperatures at the CPT exert significant control on stratospheric water vapour concentrations. Methane oxidation also contributes to water vapour concentrations in the stratosphere and plays an increasingly important role with height (Noël *et al*., 2018). It has been shown that the radiative balance of the troposphere is particularly sensitive to the concentration of water vapour in the lower stratosphere (Forster and Shine, 2002), even though these concentrations are very small relative to those in the troposphere itself. In addition, these concentrations are important for aspects of the chemistry of stratospheric ozone. The potential climate sensitivity to lower stratospheric water vapour and its role in stratospheric chemistry and radiative balance highlights the importance of understanding the vertical transport of water vapour and the accompanying dehydration processes in the TTL.

The important role of the tropical CPT in setting stratospheric water vapour concentrations was identified in a classic paper by Brewer (1949). Since then, there has been much debate about the detail of the processes, in what we now recognise as the TTL, which determine the precise concentrations. A previous summary of the state of understanding is given by Randel and

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Jensen (2013). Observations have shown clearly how variations in the CPT, leading to variations in the 'entry-value' of stratospheric water vapour concentrations, are propagated into the stratosphere by transport and mixing effects of the BDC. This is well illustrated in Figure 2, which shows the time–height variation of water vapour anomalies in the tropical stratosphere; seasonal variation in CPT temperatures imprints on water vapour concentration and slow upwelling by the BDC lifts the anomalies through the stratosphere over time, resulting in an 'atmospheric tape recorder' (Mote *et al.*, 1996).

One viewpoint (e.g. Fueglistaler *et al.*, 2005), which appears to be consistent with the observed 'tape recorder' signal, is that, to first order, concentrations are determined by temperatures on a relatively large scale, accounting for the fact that as air parcels undergo slow ascent through the region of the CPT, they sample significant geographical and temporal variability in temperatures. This variability is important in reducing water vapour concentrations below what would be expected from time-averaged, for example monthly mean, CPT temperatures (Dessler, 1998). To improve on this first order picture further processes must be invoked, some operating at relatively small scales. One is the microphysics of dehydration, for example, particle formation and sedimentation.

Another is vertical transport through the TTL via convective overshooting, when particularly strong thunderstorm complexes penetrate deep into the TTL on the timescale of a few hours. This provides a secondary pathway into and through the TTL, alongside slow radiative ascent. Certainly, rapid transport of air from the lower troposphere deep into the TTL and, more rarely, directly into the TLS via convective penetration allows for injection of very short-lived species affecting ozone concentration, such as bromine (Keeble *et al.*, 2021). Without rapid convective transport these species would otherwise play little role in stratospheric chemistry, given their short lifetimes (Robinson and Sherwood, 2006; Hosking *et al.*, 2010). Convective transport also potentially allows the cold point constraint on water vapour concentrations to be avoided. This can result in increases in water vapour concentrations in the main body of the stratosphere (Jensen *et al.*, 2007) as the additional moisture above the tropical CPT can be transported vertically by the BDC, as illustrated by the 'tape recorder' effect seen in Figure 2, and quasi-horizontally into the extratropics as illustrated in Figure 3.

The influence of convective hydration on moisture in the TTL and TLS remains an active area of research. Convective overshoots penetrate into the TTL on timescales of the order of tens of minutes and have horizontal length scales on the order of kilometres. The combination of these small-scale flows with the large-scale quasi-horizontal motions of the TTL, and the complex microphysics of moisture transport, makes modelling of convective overshoots a significant challenge. Overshoots are too small to be resolved by global climate models, which instead represent their effects with convective parameterisations. There is poor agreement between stratospheric water vapour content predicted by current global climate models and observations (Keeble *et al.*, 2021), with a general tendency for models to underpredict relative to observations, but large variations across models. The extent to which this is caused by poor representation, or indeed neglect, of convective penetration is not clear. Alongside this there have been very different approaches to estimating the quantitative contribution of convective hydration to TTL and TLS water vapour (Ueyama *et al.*, 2018; Dauhut and Hohenegger, 2022), making it difficult to interpret the differences between such estimates.

In this paper, we review current understanding of the influence of convective overshooting on moisture in the TTL and TLS. In particular, we highlight studies that have progressed understanding of this transport since the review by Randel and Jensen (2013). First, we discuss the observational evidence for convective hydration, building upon the review by Jensen *et al.* (2020). We then explore the mechanisms that lead to hydration of the TTL and subsequent transport which ultimately can hydrate the TLS and the overall influence of convective overshooting. Finally, we explore open questions and the outlook for future research on this problem and summarise our review.

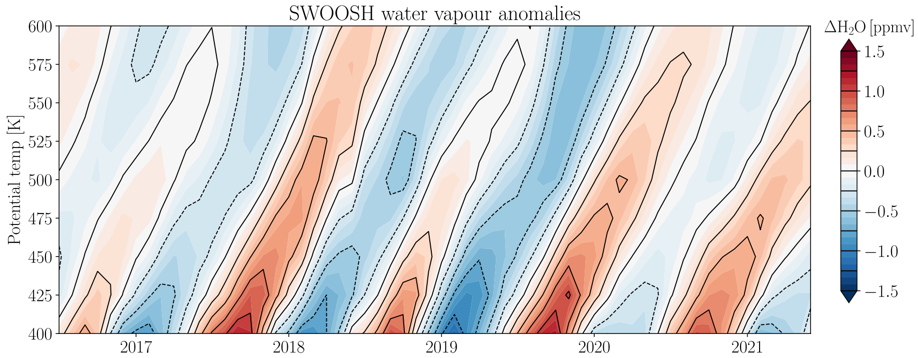

Figure 2. Atmospheric tape recorder. Zonal mean water vapour anomaly in the TLS from the SWOOSH dataset (Davis et al., *2016a) plotted on isentropic levels from 400K (around 80hPa, 18km) to 600K (around 30hPa, 24km), averaged over 15°N–15°S and with respect to the time mean from January 2006 to July 2024.*

Observational evidence for convective hydration

There is plentiful evidence for convective hydration of the TTL from *in situ* aircraft and balloon observational campaigns, as well as remote observations from satellite instruments, which capture the gradual uplift of water vapour anomalies into the TLS. Satellite observations are constrained by different limitations on temporal and spatial resolution depending on their orbit (e.g. geostationary, sun synchronous and polar) and the imaging method used (e.g. nadir- and limb-sounding – see Millán *et al.*, 2016). For example, orbits which sample a fixed point on Earth's surface only twice per day can miss the diurnal peak in convection, affecting estimates of convective activity. Satellite observations of cloud-top brightness temperatures can be used to assess the frequency and geographical distribution of overshooting convection, though uncertainties in the cloud-top altitude of up to 1km are present (Pfister *et al.*, 2022). Liu and Zipser (2005) found that 1% of tropical convective systems reach the lower TTL, and in around 0.1% of these systems, overshoots cross the CPT. In extreme cases, primarily in mesoscale convective systems (Rossow and Pearl, 2007), overshoots can penetrate directly into the TLS. The global coverage of these convective penetration events is highly spatially inhomogeneous (Nugent and Bretherton, 2023). Figure 4 shows the frequency of convection overshooting the CPT in 8 tropical regions in austral and boreal summer 2007–2010 analysed by Nugent and Bretherton (2023). Their analysis assessed the full diurnal cycle using geostationary satellite data and showed that overshoots cross the CPT 30%–40% more often over 'convectively active' land areas compared with warm ocean regions.

In situ observations from aircraft and balloons are more sparse than satellite observations but offer much finer temporal and spatial resolution, which allows direct measurements of moistening in the TTL and TLS. Several campaigns have observed convective hydration events in the tropics, such as aircraft observations from StratoClim

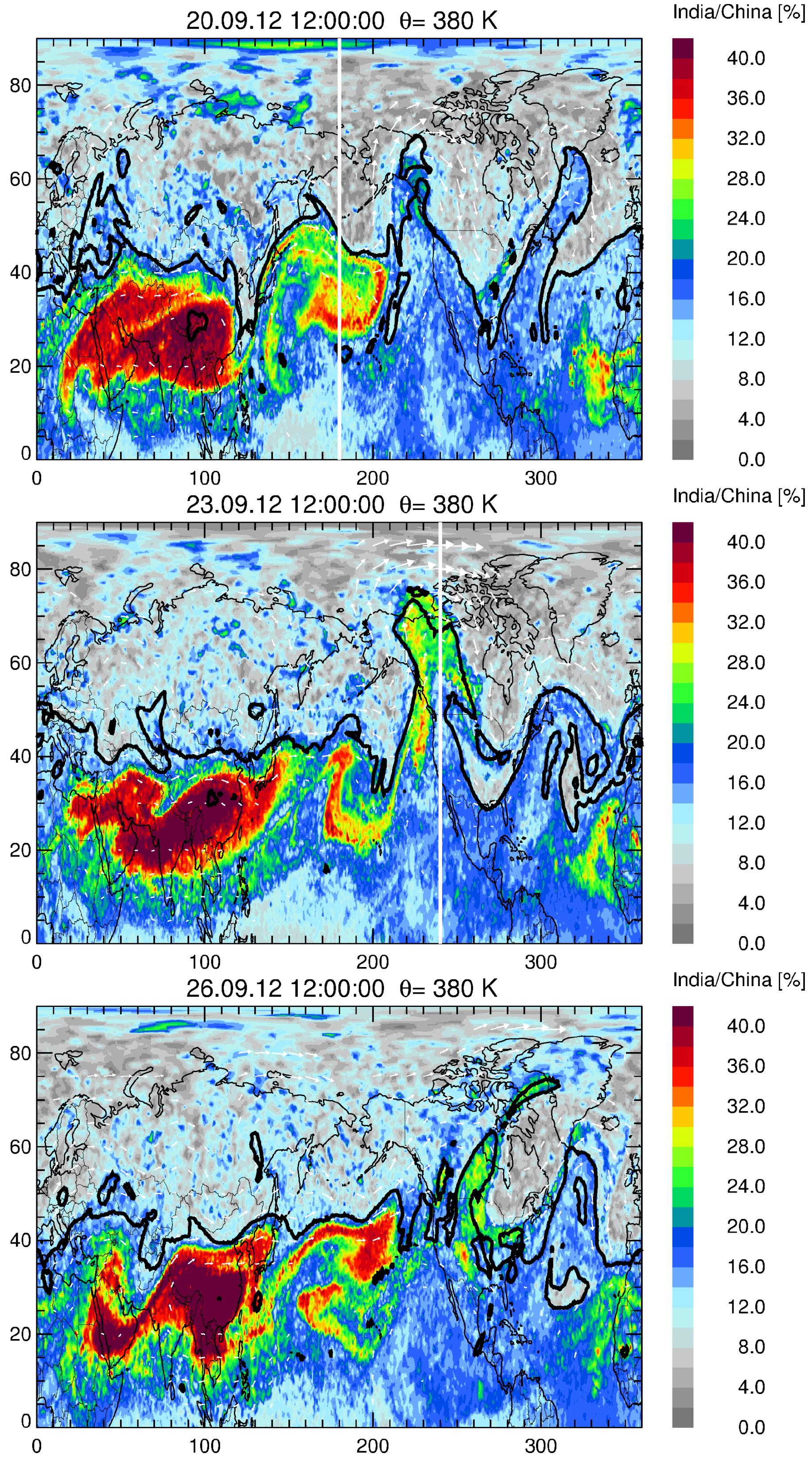

Figure 3. Horizontal distribution on the 380K surface of a modelled artificial tracer originating at the surface in the India/China region. This tracer is transported vertically by convection to give an effective localised source in the TTL in the Asian Summer Monsoon, and then horizontally both within the tropics and into the extratropical lower stratosphere. A moisture anomaly arising in the same region, through variation in CPT temperatures or convective hydration, would be correspondingly transported and therefore affect overall water vapour concentrations in the stratosphere. Snapshots are shown on 20, 23 and 26 September 2012. The horizontal winds are indicated by white arrows. The 7.2PVU surface is shown as a thick black line indicating a climatological isentropic transport barrier at 380K identified by Kunz et al. *(2015). Adapted from Vogel* et al. *(2016).*

(Bucci *et al.*, 2020), SCOUT-O3 (Vaughan *et al.*, 2008), SCOUT-AMMA (Liu *et al.*, 2010) and balloon observations from TRO-Pico (Behera *et al.*, 2022). Jensen *et al.* (2020) provide the most extensive review of observational evidence for convective hydration to date, assessing several hundred *in situ* measurements. While convective hydration above the tropical CPT occurs rarely, consistent with the aforementioned studies of cloud-top datasets, water vapour anomalies can be substantial. For example, a SCOUT-O3 flight over Brazil in austral summer 2005 observed water vapour concentrations of ~50ppmv around 400K (~17km) compared with typical values at these levels of just a few ppmv (Corti *et al.*, 2008).

The Asian Summer Monsoon (ASM) is a particularly interesting system, with potential for significant influence due to the presence of frequent tropical convection and a strong anticyclonic circulation that extends into the extratropics (Singh *et al.*, 2021). Lee *et al.* (2019) detail the evolution of a moist region observed in the StratoClim campaign during the ASM. The water vapour anomaly was formed by overshooting convection reaching altitudes up to 19km, forming a moist anomaly in the TLS up to 20km and persisting for several days. Khaykin *et al.* (2022) also consider data from the StratoClim campaign, identifying hydrated regions in the TLS which persist for several weeks.

Mechanism of moisture transport by convective overshoots

Recent modelling studies, in combination with *in situ* and remote observations, have progressed our understanding of the mechanisms that lead to vertical transport of water vapour into the TTL and ultimately the TLS. Dauhut *et al.* (2018) studied numerical simulations of 'Hector the convector', a large thunderstorm complex that forms frequently in austral summer near Darwin, Australia. Over the 10h simulation, 1500 overshoots were identified, of which around 20 penetrate deep into the TTL. Their study focuses on the processes occurring as these overshoots penetrate and collapse, on the timescale of several minutes. The aforementioned study by Lee *et al.* (2019) considers the evolution of a moist anomaly long after the overshooting convection has subsided; processes on these longer timescales can strongly modify the amount of water vapour that ultimately reaches the TLS. Here, we summarise the present understanding of the hydration mechanism, which is also illustrated in Figure 5.

As convective clouds rise through the troposphere and penetrate into the TTL, air parcels are adiabatically cooled, which

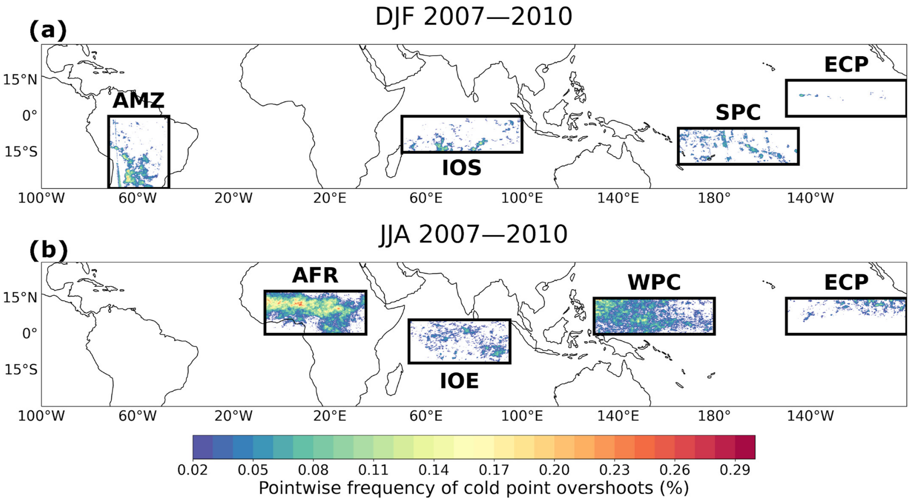

Figure 4. Frequency of convective overshoots crossing the CPT in (a) the Amazon (AMZ), Indian Ocean (IOS), South Pacific Convergence Zone (SPC) and East-Central Pacific (ECP) in boreal winter and (b) Africa (AFR), equatorial Indian Ocean (IOE), West Pacific (WPC) and ECP in boreal summer. Within the black boxes, cells are coloured white where the pointwise frequency is less than 0.02%. Reprinted from Nugent and Bretherton (2023).

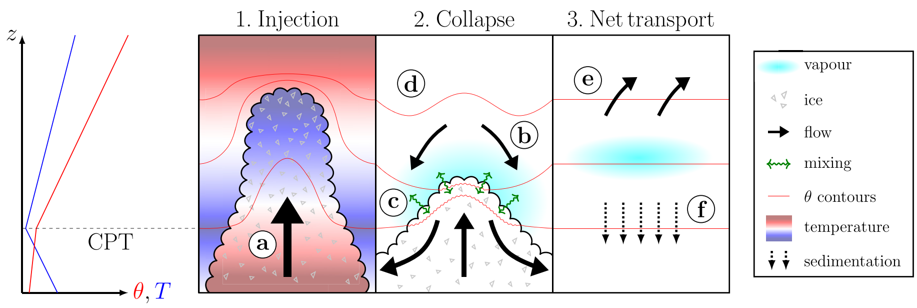

Figure 5. Schematic illustration of the mechanism for convective hydration of the TTL. Axes on the left show representative profiles of the environmental temperature T and potential temperature θ profile in the TTL, as well as the cold-point tropopause (CPT). Panel 1 shows penetration of a convective overshoot deep into the TTL, with (a) a strong updraft and significant ice loading. Shading indicates temperature. Red lines are contours of potential temperature θ. In panel 2 the overshoot collapses, causing (b) air to subside around the overshoot, enhancing (c) mixing between cold tropospheric air at the top of the overshoot and surrounding warm TTL air. Ice sublimates in the warmer air, producing a water vapour anomaly (indicated by light blue shading). (d) Gravity wave breaking may further promote mixing and associated vertical displacements can loft water vapour deeper into the TTL. Panel 3 shows that net transport of water vapour into the TLS occurs due to (e) large-scale uplift, but can be reduced by (f) the formation and sedimentation of ice.

reduces the saturation vapour concentration. Excess water vapour condenses into droplets, which eventually freeze into ice and other frozen hydrometeor species. For an individual overshoot penetrating deep into the TTL, the ice concentration is typically 100–1000 times larger than the water vapour concentration (Dauhut *et al*., 2018). In particular, overshoots which cross the CPT are significantly colder and more ice-rich than the surrounding dry (i.e. subsaturated) TTL environment, which warms with height above the CPT. Owing to the large temperature difference, the overshoot becomes negatively buoyant and collapses, generating shear between the rising updraft and subsiding flow surrounding it. Turbulence drives intense mixing between the overshoot and environment, which may be enhanced by shear within the overshoot. Large-scale wind shear in the TTL can also promote mixing. The mixing of cold and ice-rich tropospheric air with warm and dry TTL air leads to the sublimation of ice and formation of a vapour-rich pocket at the top of the overshoot. As the overshoot collapses to its level of neutral buoyancy, anomalous concentrations of water vapour remain higher in the TTL.

The hydration mechanism relies on the entrainment of subsaturated air from the TTL into the overshoot so that sublimated ice can increase the moisture content. However, when air in the TTL is supersaturated, the overshooting process can result in dehydration via 'ice scavenging' (Hassim and Lane, 2010; Khaykin *et al*., 2022), where excess vapour condenses onto convectively-lofted ice particles which grow and sediment out of the TTL. Ice scavenging drives the TTL environment back towards saturation, producing a net dehydration of the TTL. This tends to occur in overshoots that do not penetrate far into the TTL; relative humidity is typically low in the upper TTL and TLS so the rare occasions where deep convection penetrates well above the CPT almost always result in hydration (Jensen *et al*., 2020). However, Khaykin *et al*. (2022) briefly note a single flight in the StratoClim campaign during the ASM where supersaturated regions were observed above the CPT that were dehydrated by convective penetration events.

Notwithstanding the fact that moistening of the TTL occurs for overshoots that cross the CPT but do not reach the TLS, the amount of water vapour that is irreversibly transported into the TLS strongly depends on processes on longer timescales, from hours to days, during slow quasi-horizontal ascent through the TTL. During transit, gravity wave-associated temperature perturbations can temporarily cool anomalously moist regions, leading to the formation of ice which may sediment out of the TTL and reduce the net hydration of the TLS (e.g. Wright *et al*., 2011; Tissier and Legras, 2016; Pan *et al*., 2019).

Convective overshoots generate gravity waves which propagate outwards and upwards, producing remote effects as well as local influence on the overshoot (e.g. Hassim and Lane, 2010; Sang *et al*., 2018; Lee *et al*., 2019). These waves have associated temperature and vertical velocity perturbations which propagate through the TTL and become amplified when waves break in the presence of vertical wind shear. Sang *et al*. (2018) found that gravity wave breaking promotes mixing at the top of the overshoot, thereby increasing the net hydration of the TTL by entraining larger volumes of warm stratospheric air, which then sublimates greater amounts of ice. It was also found that the presence of large-scale vertical shear near the CPT limits the amplitude of gravity waves and reduces the overall transport as a result.

The remote effects of gravity waves have been invoked as an explanation for the 'jumping cirrus' phenomenon, where hydration occurs up to 1–3km above the overshoot (Wang, 2003; Iwasaki and Yamaguchi, 2022). The mechanism was examined by Hassim and Lane (2010) using numerical simulations which showed that upward displacement of TTL air by breaking gravity waves results in ice formation by adiabatic cooling, which is then mixed into the subsaturated (and warmer) environment of the upper TTL and TLS. However, it is unclear whether the large air parcel displacements are solely associated with gravity wave breaking – which necessitates sufficient vertical wind shear in the TTL environment – or associated with motion of the

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overshoot itself (Frey *et al*., 2015). Figure 6 shows an example of our own numerical simulations of convective penetration of a stably stratified layer (Powell *et al*., 2024), which may be used to explore the problem further using an idealised representation of the flow.

The TTL temperature structure plays an important role in determining the hydration potential of an overshoot. A cooler TTL environment limits the amount of sublimated ice that can remain after convective injection, since the saturation vapour concentration is lower. The increase of TTL temperatures with height above the CPT makes the maximum penetration height of a convective overshoot a key factor (Dauhut *et al*., 2018). A larger maximum height allows mixing of even warmer air from the upper TTL into the overshoot, increasing the saturation concentration and allowing more ice particles to sublimate. This supports an earlier study by Sherwood and Dessler (2001) in which an idealised convective overshooting scheme coupled to a model of the large-scale circulation showed particular sensitivity to the distribution of heights where mixing with air in the TTL occurs. Numerical simulations and idealised models remain a fruitful avenue for studying the competition between processes such as mixing, shear and sedimentation in convective penetration.

Influence on stratospheric water vapour budget

It is understood that water vapour content in the TLS is primarily governed by temperatures at the CPT, as is evident from the tape recorder where the seasonal cycle imprints on observations. On shorter timescales and in localised regions, convection can exert an influence on moisture which is regarded as second order (Schiller, 2009; Wright *et al*., 2011). Estimates of the convective contribution to water vapour mass input into the TLS generally lie in the 10%–20% range, although some estimates are significantly lower than this; Table 1 summarises recent estimates. The methodology can broadly be categorised into those using Lagrangian trajectory models and global storm-resolving models.

There are differences in the way that convective 'contribution' is quantified which makes comparison difficult. For example, Schoeberl *et al*. (2018) and Ueyama *et al*. (2015, 2018) compare predicted water vapour concentrations entering the stratosphere that would be estimated with and without convection, while Dauhut and Hohenegger (2022) consider the TLS water vapour budget directly from a global storm-resolving model. Estimates are also calculated with varying spatial and temporal domains which might not be expected to agree given the seasonal variation in convective activity and TTL temperatures. Estimates are also sensitive to the choice of cloud microphysics scheme, since the potential for an overshoot to hydrate the TTL relies on sublimation of convectively-lofted ice. Similarly, dehydration (by overshooting and also in subsequent quasi-horizontal transit through the TTL) relies on an appropriate representation of the growth of ice particles in a supersaturated and very cold environment (Jensen and Pfister, 2004; Randel and Jensen, 2013).

Lagrangian models calculate trajectories using reanalysis datasets and estimate water vapour transport using a microphysical scheme, some of which are simplified (e.g. Schoeberl *et al*., 2014, 2018) and some which

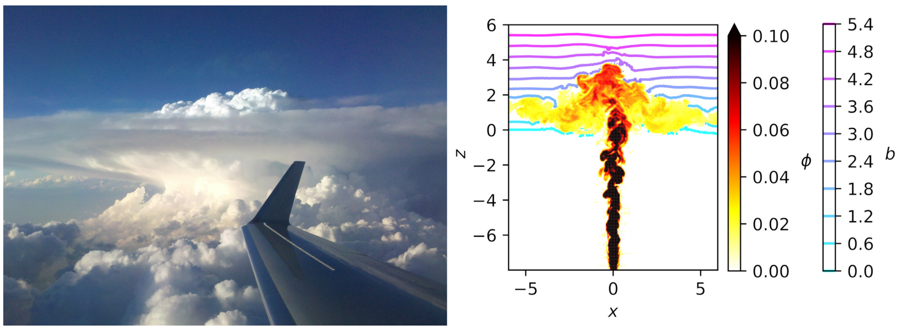

Figure 6. Left: deep convection with overshooting tops over North America, 12 May 2012, photographed from an aircraft. Photograph courtesy of NASA Earth Observatory. Right: numerical simulation of a buoyant plume penetrating into a stably stratified layer carrying a passive tracer from Powell et al. *(2024). Contours of buoyancy* b *(linearly related to potential temperature* θ*) shown with passive tracer concentration* ϕ *overlaid.*

Table 1

Estimates of the contribution of convective hydration to the TTL water vapour budget.

Methodology	*Study*	*Period and domain*	*Contribution definition*	*Estimated contribution (%)*
Lagrangian trajectory model with simple cloud model	Schoeberl *et al.* (2014)	Boreal winter 2008/2009, 60°S to 60°N	Water vapour mass input at 19km	13
	Schoeberl *et al.* (2018)	Boreal winter 2008/2009, 30°S to 30°N	Net change in water vapour at last convective encounter, with and without convective transport	2
Lagrangian trajectory model with expensive microphysics scheme	Ueyama *et al.* (2015)	7-day period in boreal summer 2007, 20°S to 20°N	Net change in water vapour at 100hPa with and without convection	14
	Ueyama *et al.* (2018)	7-day period in boreal winter 2006/2007, 10°S to 50°N		15
Simulation of a single convective system upscaled to tropics	Dauhut *et al.* (2015)	Full year extrapolated from event in boreal winter 2005, 20°S to 20°N	Water vapour mass input across 100hPa	18
Global storm-resolving model	Dauhut and Hohenegger (2022)	40-day period in boreal summer 2016, 10°S to 30°N	Water vapour mass input to 17–20km altitude	11

Note: The different methodologies and definitions of convective contribution makes comparison between studies difficult. Other than Schoeberl *et al.* (2018), these estimates support the conclusion that convective overshooting makes a non-negligible but modest contribution.

are more sophisticated and computationally expensive (e.g. Ueyama *et al*., 2015, 2018). Estimates are limited by the accuracy and resolution of temperature and wind fields used to calculate trajectories, as well as the need to model convective encounters, which relies on the resolution and accuracy of cloud models and datasets. It should also be noted that estimates by Ueyama *et al*. (2015, 2018) are calculated slightly below the CPT rather than deep in the TTL, making them difficult to compare with other studies; the contribution may be underestimated since the lower part of the TTL tends to be close to saturation, or overestimated as the number of convective systems that actually penetrate the CPT is much smaller.

Global storm-resolving models are capable of resolving individual convective systems and therefore do not rely on parameterisations of convection. Other uncertainties remain, as individual convective overshoots are not fully resolved. Dauhut and Hohenegger (2022) present the first estimate of the vertical water vapour mass flux into the TLS using a global storm-resolving model, with results that seem to be broadly in agreement with many of the trajectory-based estimates. Their study highlights the difficulty in estimating the convective contribution from water vapour anomalies above deep convection, as seen in Jensen *et al*. (2020). Strong horizontal divergence above convective systems rapidly decreases water vapour anomalies so that measurements taken shortly after convective injection result in an underestimation of the net moisture transport.

Open questions and outlook

Most global climate models predict that stratospheric water vapour content will increase under increased greenhouse gas forcing (Keeble *et al*., 2021). The future role of convective hydration under climate change is an important and complex question, not least because of the feedbacks between stratospheric water vapour and global surface temperatures (Solomon *et al*., 2010). Understanding the response to climate forcing is particularly difficult, owing to the coarse resolution of global climate models. Homeyer (2015) found that storm altitudes and water vapour transport are sensitive to spatial resolution in models; storm intensity increases with finer horizontal resolution (increasing transport) while the tropopause becomes sharper with increasing vertical resolution (decreasing transport).

Studies have shown potential changes in the geographical distribution and frequency of overshooting convection in response to a warming atmosphere, with a more equal balance between convection over land and the ocean, and an increase in water vapour transport in the TTL. Smith *et al*. (2022) study Lagrangian trajectories calculated using output from a single climate model, with water vapour content estimated based on the coldest temperatures encountered along each trajectory. The analysis suggests that in response to increases in atmospheric CO_2, the CPT moves upwards and increases in temperature, reducing the CPT temperature constraint on water vapour concentration. Convection reaches a higher altitude but does not penetrate further above the raised CPT. While the number of convective overshoots increases, the proportion of overshoots penetrating deep into the TTL does not. These results imply a greater water vapour mass input to the TTL, but Smith *et al*. (2022) note that the relative contribution of convective hydration to the water vapour budget remains roughly constant due to the proportional increase in CPT temperatures. Wu *et al*. (2023) find consistent results in their study of the frequency of overshooting convection in a global storm-resolving model, particularly noting that the frequency of overshoots increases over oceans and decreases over land.

Summary

Understanding the water vapour budget of the TLS and the response to climate forcing is crucial, given its impacts on radiative balance and stratospheric chemistry. While deep convective systems penetrate the TTL relatively infrequently, some of these overshoots cross the cold-point tropopause, which allows rapid convective transport to avoid the cold-point constraint on water vapour concentrations. In these overshoots, mixing of moist, cold tropospheric air with warm, dry stratospheric air results in a net hydration of the TTL. Overshoots penetrating into supersaturated regions of the TTL can instead result in a net dehydration. Following transient convective injection events, which last on the order of 10min, slow quasi-horizontal upwelling lifts water vapour anomalies into the TLS over several weeks, during which gravity wave perturbations (among other processes) can significantly modify the amount of water vapour that ultimately reaches the TLS.

The process of convective hydration of the TTL and irreversible transport into the TLS is complex. Determining net hydration or dehydration of the TTL is often an oversimplification; some layers of the TTL may be hydrated by a convective overshoot while other layers are dehydrated, and water vapour already present in the TTL can be redistributed. Hydration of the TTL by convective overshoots is a highly localised process, occurring primarily in mesoscale convective systems of a few kilometres and with strong sensitivity to the maximum altitude reached by overshooting cloud tops.

While seasonal and interannual variation of temperatures at the cold-point tropopause primarily control water vapour concentration in the TLS, recent estimates suggest there is a modest influence of convective hydration on TLS water vapour. However, other recent studies suggest this contribution as a fraction of total concentrations is not expected to increase under climate change.

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Author contributions

C. W. Powell: Conceptualization; formal analysis; investigation; project administration; writing – review and editing; writing – original draft; visualization. **P. H. Haynes:** Writing – review and editing; conceptualization; investigation. **A. D. Ming:** Writing – review and editing; conceptualization; formal analysis; visualization. **J. R. Taylor:** Writing – review and editing.

Data availability statement

SWOOSH data can be downloaded from the National Oceanic and Atmospheric Administration website (Davis *et al*., 2016b).

References

Behera AK, **Rivière ED**, **Khaykin SM** *et al*. 2022. On the cross-tropopause transport of water by tropical convective overshoots: a mesoscale modelling study constrained by in situ observations during the TRO-Pico field campaign in Brazil. *Atmos. Chem. Phys.* **22**(2): 881–901. https://doi.org/10.5194/acp-22-881-2022

Brewer AW. 1949. Evidence for a world circulation provided by the measurements of helium and water vapour distribution in the stratosphere. *Q. J. R. Meteorol. Soc.* **75**(326): 351–363. https://doi.org/10.1002/qj.49707532603

Bucci S, **Legras B**, **Sellitto P** *et al*. 2020. Deep-convective influence on the upper troposphere–lower stratosphere composition in the Asian monsoon anticyclone region: 2017 StratoClim campaign results. *Atmos. Chem. Phys.* **20**(20): 12193–12210. https://doi.org/10.5194/acp-20-12193-2020

Butchart N. 2014. The Brewer-Dobson circulation. *Rev. Geophys.* **52**: 157–184. https://doi.org/10.1002/2013RG000448

Corti T, **Luo BP**, **De Reus M** *et al*. 2008. Unprecedented evidence for deep convection hydrating the tropical stratosphere. *Geophys. Res. Lett.* **35**(10): 2008GL033641. https://doi.org/10.1029/2008GL033641

Dauhut T, **Chaboureau J-P**, **Escobar J** *et al*. 2015. Large-eddy simulations of Hector the convector making the stratosphere wetter. *Atmos. Sci. Lett.* **16**(2): 135–140. https://doi.org/10.1002/asl2.534

Dauhut T, **Chaboureau J-P**, **Haynes P** *et al*. 2018. The mechanisms leading to a stratospheric hydration by overshooting convection. *J. Atmos. Sci.* **75**(12): 4383–4398. https://doi.org/10.1175/JAS-D-18-0176.1

Dauhut T, **Hohenegger C**. 2022. The contribution of convection to the stratospheric water vapor: the first budget using a global storm-resolving model. *J. Geophys. Res. Atmos.* **127**(5): e2021JD036295. https://doi.org/10.1029/2021JD036295

Davis SM, **Rosenlof KH**, **Hassler B** *et al*. 2016a. The stratospheric water and ozone satellite homogenized (SWOOSH) database: a long-term database for climate studies. *Earth Syst. Sci. Data* **8**(2): 461–490.

Davis SM, **Rosenlof KH**, **Hassler B** *et al*. 2016b. The Stratospheric Water and Ozone Satellite Homogenized (Swoosh) Data. NOAA: US https://csl.noaa.gov/groups/csl8/swoosh/ [accessed January 2024].

Dessler AE. 1998. A reexamination of the "stratospheric fountain" hypothesis. *Geophys. Res. Lett.* **25**(22): 4165–4168. https://doi.org/10.1029/1998GL900120

>**Forster PMDF**, **Shine KP**. 2002. Assessing the climate impact of trends in stratospheric water vapor. *Geophys. Res. Lett.* **29**(6): 10-1–10-4. https://doi.org/10.1029/2001GL013909

Frey W, **Schofield R**, **Hoor P** *et al*. 2015. The impact of overshooting deep convection on local transport and mixing in the tropical upper troposphere/lower stratosphere (UTLS). *Atmos. Chem. Phys.* **15**(11): 6467–6486. https://doi.org/10.5194/acp-15-6467-2015

Fueglistaler S, **Bonazzola M**, **Haynes PH** *et al*. 2005. Stratospheric water vapor predicted from the Lagrangian temperature history of air entering the stratosphere in the tropics. *J. Geophys. Res.* **110**(D8): 2004JD005516. https://doi.org/10.1029/2004JD005516

Fueglistaler S, **Dessler AE**, **Dunkerton TJ** *et al*. 2009. Tropical tropopause layer. *Rev. Geophys.* **47**(1): 2008RG000267. https://doi.org/10.1029/2008RG000267

Hassim MEE, **Lane TP**. 2010. A model study on the influence of overshooting convection on TTL water vapour. *Atmos. Chem. Phys.* **10**(20): 9833–9849. https://doi.org/10.5194/acp-10-9833-2010

Homeyer CR. 2015. Numerical simulations of extratropical tropopause-penetrating convection: sensitivities to grid resolution. *J. Geophys. Res. Atmos.* **120**(14): 7174–7188. https://doi.org/10.1002/2015JD023356

Hosking JS, **Russo MR**, **Braesicke P** *et al*. 2010. Modelling deep convection and its impacts on the tropical tropopause layer. *Atmos. Chem. Phys.* **10**(22): 11175–11188. https://doi.org/10.5194/acp-10-11175-2010

Iwasaki S, **Yamaguchi T**. 2022. Characteristics of a jumping cirrus retrieved from rapid scans by Himawari-8. *J. Geophys. Res. Atmos.* **127**(22): e2022JD037539. https://doi.org/10.1029/2022JD037539

Jensen EJ, **Ackerman AS**, **Smith JA**. 2007. Can overshooting convection dehydrate the tropical tropopause layer? *J. Geophys. Res.* **112**(D11): 2006JD007943. https://doi.org/10.1029/2006JD007943

Jensen EJ, **Pan LL**, **Honomichl S** *et al*. 2020. Assessment of observational evidence for direct convective hydration of the lower stratosphere. *J. Geophys. Res. Atmos.* **125**(15): e2020JD032793. https://doi.org/10.1029/2020JD032793

Jensen EJ, **Pfister L**. 2004. Transport and freeze-drying in the tropical tropopause layer. *J. Geophys. Res.* **109**(D2): 2003JD004022. https://doi.org/10.1029/2003JD004022

Keeble J, **Hassler B**, **Banerjee A** *et al*. 2021. Evaluating stratospheric ozone and water vapour changes in CMIP6 models from 1850 to 2100. *Atmos. Chem. Phys.* **21**(6): 5015–5061. https://doi.org/10.5194/acp-21-5015-2021

Khaykin SM, **Moyer E**, **Krämer M** *et al*. 2022. Persistence of moist plumes from overshooting convection in the Asian monsoon anticyclone. *Atmos. Chem. Phys.* **22**(5): 3169–3189. https://doi.org/10.5194/acp-22-3169-2022

Kunz A, **Sprenger M**, **Wernli H**. 2015. Climatology of potential vorticity streamers and associated isentropic transport pathways across PV gradient barriers. *J. Geophys. Res. Atmos.* **120**(9): 3802–3821. https://doi.org/10.1002/2014JD022615

Lee K-O, **Dauhut T**, **Chaboureau J-P** *et al*. 2019. Convective hydration in the tropical tropopause layer during the StratoClim aircraft campaign: pathway of an observed hydration patch. *Atmos. Chem. Phys.* **19**(18): 11803–11820. https://doi.org/10.5194/acp-19-11803-2019

Liu C, **Zipser EJ**. 2005. Global distribution of convection penetrating the tropical tropopause. *J. Geophys. Res.* **110**(D23): 2005JD006063. https://doi.org/10.1029/2005JD006063

Liu XM, **Rivière ED**, **Marécal V** *et al*. 2010. Stratospheric water vapour budget and convection overshooting the tropopause: modelling study from SCOUT-AMMA. *Atmos. Chem. Phys.* **10**(17): 8267–8286. https://doi.org/10.5194/acp-10-8267-2010

Millán LF, **Livesey NJ**, **Santee ML** *et al*. 2016. Case studies of the impact of orbital sampling on stratospheric trend detection and derivation of tropical vertical velocities: solar occultation vs. limb emission sounding. *Atmos. Chem. Phys.* **16**(18): 11521–11534. https://doi.org/10.5194/acp-16-11521-2016

Mote PW, **Rosenlof KH**, **McIntyre ME** *et al*. 1996. An atmospheric tape recorder: the imprint of tropical tropopause temperatures on stratospheric water vapor. *J. Geophys. Res. Atmos.* **101**(D2): 3989–4006. https://doi.org/10.1029/95JD03422

Noël S, **Weigel K**, **Bramstedt K** *et al*. 2018. Water vapour and methane coupling in the stratosphere observed using SCIAMACHY solar occultation measurements. *Atmos. Chem. Phys.* **18**(7): 4463–4476. https://doi.org/10.5194/acp-18-4463-2018

Nugent JM, **Bretherton CS**. 2023. Tropical convection overshoots the cold point tropopause nearly as often over warm oceans as over land. *Geophys. Res. Lett.* **50**(21): e2023GL105083. https://doi.org/10.1029/2023GL105083

Pan LL, **Honomichl SB**, **Thornberry T** *et al*. 2019. Observational evidence of horizontal transport-driven dehydration in the TTL. *Geophys. Res. Lett.* **46**(13): 7848–7856. https://doi.org/10.1029/2019GL083647

Pfister L, **Ueyama R**, **Jensen EJ** *et al*. 2022. Deep convective cloud top altitudes at high temporal and spatial resolution. *Earth Space Sci.* **9**(11): e2022EA002475. https://doi.org/10.1029/2022EA002475

Powell CW, **Haynes PH**, **Taylor JR**. 2024. Diagnosing tracer transport in convective penetration of a stably stratified layer. *J. Fluid Mech.* **997**: A48. https://doi.org/10.1017/jfm.2024.662

Randel WJ, **Jensen EJ**. 2013. Physical processes in the tropical tropopause layer and their roles in a changing climate. *Nat. Geosci.* **6**(3): 169–176. https://doi.org/10.1038/ngeo1733

Robinson FJ, **Sherwood SC**. 2006. Modeling the impact of convective entrainment on the tropical tropopause. *J. Atmos. Sci.* **63**(3): 1013–1027. https://doi.org/10.1175/JAS3673.1

Rossow WB, **Pearl C**. 2007. 22-Year survey of tropical convection penetrating into the lower stratosphere. *Geophys. Res. Lett.* **34**(4): 2006GL028635. https://doi.org/10.1029/2006GL028635

Sang W, **Huang Q**, **Tian W** *et al*. 2018. A large eddy model study on the effect of overshooting convection on lower stratospheric water vapor. *J. Geophys. Res. Atmos.* **123**(18): 10023–10036. https://doi.org/10.1029/2017JD028069

Schiller C. 2009. Hydration and dehydration at the tropical tropopause. *Atmos. Chem. Phys.* **9**(24): 9647–9660. https://doi.org/10.5194/acp-9-9647-2009

Schoeberl MR, **Dessler AE**, **Wang T** *et al*. 2014. Cloud formation, convection, and stratospheric dehydration. *Earth Space Sci.* **1**(1): 1–17. https://doi.org/10.1002/2014EA000014

Schoeberl MR, **Jensen EJ**, **Pfister L** *et al*. 2018. Convective hydration of the upper troposphere and lower stratosphere. *J. Geophys. Res. Atmos.* **123**(9): 4583–4593. https://doi.org/10.1029/2018JD028286

Sherwood SC, **Dessler AE**. 2001. A model for transport across the tropical tropopause. *J. Atmos. Sci.* **58**(7): 765–779.

Singh BB, **Krishnan R**, **Ayantika DC** *et al*. 2021. Linkage of water vapor distribution in the lower stratosphere to organized Asian summer monsoon convection. *Clim. Dyn.* **57**(7–8): 1709–1731. https://doi.org/10.1007/s00382-021-05772-2

Smith JW, **Bushell AC**, **Butchart N** *et al*. 2022. The effect of convective injection of ice on stratospheric water vapor in a changing climate. *Geophys. Res. Lett.* **49**(9): e2021GL097386. https://doi.org/10.1029/2021GL097386

Solomon S, **Rosenlof KH**, **Portmann RW** *et al*. 2010. Contributions of stratospheric water vapor to decadal changes in the rate of global warming. *Science* **327**(5970): 1219–1223. https://doi.org/10.1126/science.1182488

Tissier A-S, **Legras B**. 2016. Convective sources of trajectories traversing the tropical tropopause layer. *Atmos. Chem. Phys.* **16**(5): 3383–3398. https://doi.org/10.5194/acp-16-3383-2016

Ueyama R, **Jensen EJ**, **Pfister L**. 2018. Convective influence on the humidity and clouds in the tropical tropopause layer during boreal summer. *J. Geophys. Res.*

Atmos. **123**(14): 7576–7593. https://doi.org/10.1029/2018JD028674

Ueyama R, **Jensen EJ**, **Pfister L** *et al.* 2015. Dynamical, convective, and microphysical control on wintertime distributions of water vapor and clouds in the tropical tropopause layer. *J. Geophys. Res.* **120**(19): 10483–10500. https://doi.org/10.1002/2015JD023318

Vaughan G, **Schiller C**, **MacKenzie AR** *et al.* 2008. SCOUT-O3/ACTIVE: high-altitude aircraft measurements around deep tropical convection. *Bull. Amer. Meteor. Soc.* **89**(5): 647–662. https://doi.org/10.1175/BAMS-89-5-647

Vogel B, **Günther G**, **Müller R** *et al.* 2016. Long-range transport pathways of tropospheric source gases originating in Asia into the northern lower stratosphere during the Asian monsoon season 2012. *Atmos. Chem. Phys.* **16**(23): 15301–15325. https://doi.org/10.5194/acp-16-15301-2016

Wang PK. 2003. Moisture plumes above thunderstorm anvils and their contributions to cross-tropopause transport of water vapor in midlatitudes. *J. Geophys. Res. Atmos.* **108**(6): 4194. https://doi.org/10.1029/2002jd002581

Wright JS, **Fu R**, **Fueglistaler S** *et al.* 2011. The influence of summertime convection over Southeast Asia on water vapor in the tropical stratosphere. *J. Geophys. Res.* **116**(D12): D12302. https://doi.org/10.1029/2010JD015416

Wu X, **Fu Q**, **Kodama C**. 2023. Response of tropical overshooting deep convection to global warming based on global cloud-resolving model simulations. *Geophys. Res. Lett.* **50**(14): e2023GL104210. https://doi.org/10.1029/2023GL104210

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