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Physical processes in the tropical tropopause layer and their roles in a changing climate

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Tropical climate and the composition of the global upper atmosphere are affected by the tropical tropopause layer — the atmospheric transition zone between the well-mixed, convective troposphere (up to altitudes of 12–14 km) and the highly stratified stratosphere (above about 18 km). Featuring chemical and dynamical properties that are midway between those of the troposphere and stratosphere, the tropopause layer is maintained by a complex interplay between large- and small-scale circulation patterns, deep convection, clouds and radiation. Tropospheric air enters the stratosphere primarily in the tropics. Ozone- and aerosol-related constituents of the global stratosphere, as well as water vapour content, are therefore largely determined by the composition of the air near the tropical tropopause. Over the past years, it has emerged that both slow ascent and rapid deep convection contribute to the composition and thermal structure of the tropical tropopause layer. Ice formation processes at low temperatures affect the efficacy of freeze drying as air passes through the cold tropopause region. Transport and mixing in the tropopause region has been found to be closely linked with the Asian monsoon and other tropical circulation systems. Given these connections, climate change is expected to influence the tropopause layer, for example through enhanced large-scale upwelling of air and potential changes in tropical convection, air temperature, chemical composition and cirrus.

In the tropics, temperatures decrease with height from the surface up to an altitude of around 17 km, and then increase at higher altitudes in the stratosphere. The temperature minimum is termed the cold-point tropopause. The troposphere (altitudes below 17 km) is characterized by regions of persistent strong convection over continents and the western Pacific Ocean, and the overall thermal structure is mainly determined by large-scale radiative–convective equlibrium. Deep convection drives the general circulation of the tropical troposphere, and controls transport of energy and water vapour as well as other trace constituents in the climate system. Strong tropical convection, which covers approximately 10% of the tropics at any given time, is characterized by inflow at low levels, rapid vertical transport and outflow near altitudes 12–14 km (ref. 1). Less frequent intense convection can reach altitudes of the cold-point tropopause or beyond[2–4], although the contribution of such high-reaching convection to the global mass flux into the stratosphere is believed to be rather small. Thermal structure above the well-mixed troposphere is primarily determined by radiative–dynamical balance; absorption of solar ultraviolet light by stratospheric ozone accounts for the temperature increase above 17 km. The region between the top of strong convective outflow near 12–14 km and altitudes near 18 km (where the last overshooting convective towers occasionally reach) has physical and chemical characteristics midway between the troposphere and stratosphere, and is termed the tropical tropopause layer (TTL)[5] (Fig. 1a). There are alternative specific definitions of the TTL (refs 6,7), but they all capture the same key feature of transition from convective to radiative regimes.

There is a large-scale mean upward circulation within the upper TTL above approximately 15 km, which influences temperatures and composition. This upwelling is dynamically driven by large- and small-scale waves originating in the extratropics as well as the tropics (Fig. 1b). This mean circulation maintains temperatures well below radiative equilibrium, and the TTL is the coldest region in the atmosphere (below 50 km) outside of the polar winter stratosphere. The mean upward circulation is also responsible for the primary transport of air from the troposphere to the stratosphere, forming the tropical component of the mean overturning circulation in the stratosphere, termed the Brewer Dobson circulation. Hence the TTL provides the physical and chemical boundary conditions for air entering the global stratosphere. Dramatic evidence of this behaviour is provided by observations of stratospheric water vapour: air transiting the TTL is substantially dehydrated on passing the cold tropical tropopause, leading to the extreme dryness (just a few ppm of water vapour) of the global stratosphere[8]. Because tropopause temperatures have a strong annual cycle, there is a resulting seasonal cycle in stratospheric water vapour imposed near the cold point, which propagates upwards into the stratosphere with the mean upwelling circulation (the so-called stratospheric tape recorder[9]).

There is frequent quasi-horizontal two-way exchange between the TTL and the extratropical lower stratosphere in both hemispheres, influencing composition in all regions. Rapid transport from the tropics to middle latitudes occurs above the subtropical jets (Fig. 1a). This transport is especially evident in lower stratosphere water vapour (Fig. 2), as air dehydrated in the tropics during boreal winter transits to extratropics in both hemispheres[10]. Enhanced poleward transport events are often associated with intrusions of tropical air and a secondary tropopause over subtropics and middle latitudes[11,12]. Evidence of transport into the tropics is apparent in tracers with stratospheric origin (such as ozone or HCl; ref. 13) that are observed within the deep tropics; this transport seems largest in the Northern Hemisphere during boreal summer[14], probably tied to monsoonal circulations[15].

The high, cold region of the TTL has a disproportionately large influence on global climate through several processes. High-altitude thin cirrus clouds and convective anvils are ubiquitous features of the TTL (refs 16,17), with strong feedbacks on tropical

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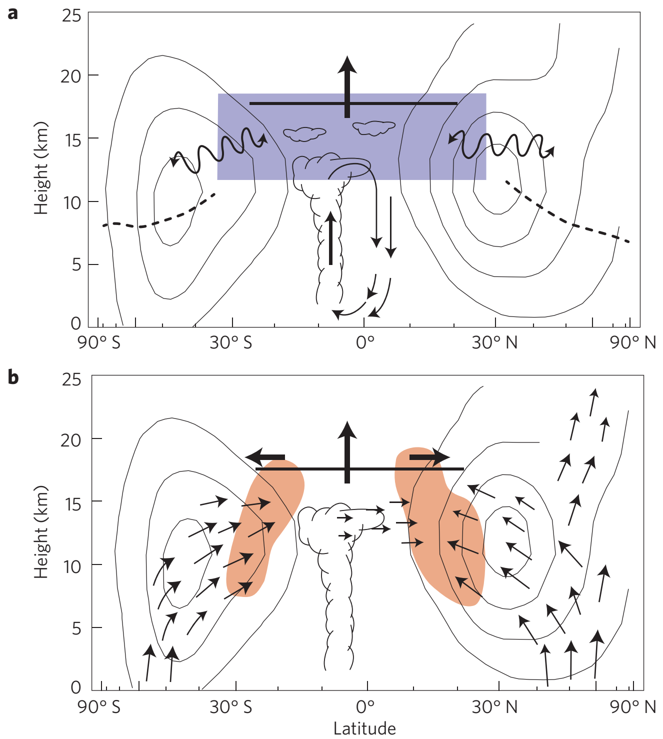

Figure 1 | Schematic of the large-scale structure and circulation of the TTL in the latitude–height plane. The influence of tropical deep convection and interactions with other regions of the atmosphere are highlighted. The solid line in the tropics near 17 km indicates the cold-point tropopause, and the dotted lines denote the extratropical tropopause. Contours represent the zonal average winds, and the wriggled lines highlight two-way transport between the TTL (blue region) and extratropics. The upward arrow across the tropical tropopause indicates the large-scale upwelling associated with the Brewer-Dobson circulation. The latitude scale is proportional to area (weighted by cos (latitude)). **b**, Large-scale dynamical structure of the zonal mean atmosphere, highlighting the propagation of tropical and extratropical waves (denoted by arrows) that dissipate in the subtropics (red regions), inducing polewards motion in the subtropical lower stratosphere and time mean upwelling within the TTL.

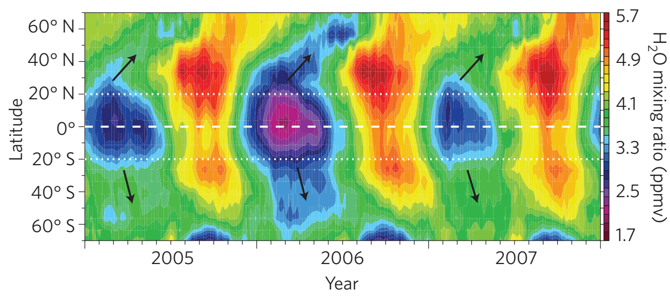

Figure 2 | Satellite observations of water vapour in the lower stratosphere demonstrate the transport of air from the TTL to high latitudes. Air is dehydrated near the cold-point tropical tropopause over latitudes 20° N–S, with a strong annual cycle related to minimum tropical temperatures during the boreal winter season. This dehydrated air is effectively transported to middle and high latitudes in both hemispheres; note that the interannual changes in tropical water vapour are reflected at higher latitudes. The boreal summer maxima are related to monsoonal circulations. These data (for the 390 K potential temperature level) are from the Aura MLS instrument.

radiative balances[18]. Near-continuous observations of cirrus clouds by space-based lidar since 2006[19] have provided improved understanding of large-scale structure[17,20] and variability, showing that TTL cirrus are closely coupled to well-known modes of tropical circulations such as the Madden-Julian oscillation and El Niño/Southern Oscillation[21,22].

Atmospheric radiative fluxes are especially sensitive to temperatures and water vapour concentrations in the tropical upper troposphere[23]. Theoretical arguments suggest that the altitude of tropical deep convective outflow is closely coupled to radiative balances in the TTL (ref. 24), so that TTL processes can influence convective outflow and the equilibrium tropospheric circulation[25]. Further, the TTL chemical composition has fundamental importance for chemical behaviour of the stratosphere, and for global ozone in particular; for example, halogen species contributing to stratospheric chemical ozone loss cycles must transit through the TTL. For these reasons, understanding the detailed circulations of the TTL and capturing the underlying physics in chemistry–climate models is of key importance. However, the complexity of the region, involving transition from convective to radiatively controlled regimes, with a combination of large- and small-scale circulations and complex cloud and radiative behaviour, makes comprehensive modelling of this region a daunting challenge.

Dynamics of tropical upwelling, the annual cycle and tracers

The large-scale mean upwelling circulation in the TTL is forced by the action of atmospheric waves originating primarily outside of the tropics, such as mid-latitude weather systems or high latitude stratospheric planetary waves. Extratropical disturbances tend to propagate towards low latitudes during their life cycles, and the dissipation of waves through mechanical or thermal damping produce mean poleward motions in the stratosphere which induce upwelling over tropical latitudes (Fig. 1b). Because of the dynamical balances inherent to large-scale atmospheric circulations, tropical upwelling for slow time scales such as the seasonal cycle or long-term interannual trends is most effectively forced by wave dissipation in the subtropics[26,27]. In contrast, faster variability such as week-to-week changes can be effectively forced by high-latitude waves. The importance of subtropical wave forcing has been emphasized in recent studies focused on the annual cycle[28], enhanced upwelling during El Niño events[29] and the response to climate change[30,31].

The annual cycle within the TTL is especially interesting, with an approximate factor-of-two variation in the strength of upwelling and temperature changes of up to 8 K in the lower stratosphere, with faster upwelling and colder temperatures during boreal winter. The annual cycle in temperatures occurs exclusively above 15 km; there is almost zero seasonal variation in tropical temperatures below this level. It was originally proposed that the tropical annual cycle was a response to the annual cycle of stratospheric planetary wave forcing in winter–spring high latitudes, especially the large Northern Hemisphere peak during December–March[32,33]. However, owing to improved dynamical understanding of slow time scales, more recent analyses of the annual cycle have focused on the importance of extratropical waves dissipating in the subtropics[34–36]. It also seems that planetary waves generated within the tropics by deep convection contribute to forcing mean upwelling[28,37–39]. However, the details of how these different mechanisms contribute to stronger upwelling during boreal winter, driving the annual cycle in temperatures and tracers, remain to be clarified.

An important consequence of the TTL upwelling is the vertical transport of trace constituents. This is especially important for ozone, which has a strong vertical gradient across the TTL owing to photochemical production in the stratosphere. The annual cycle in upwelling produces a large variation in lower stratospheric ozone, of up to a factor of two near the tropopause, which is approximately in phase with temperatures. Recent work has emphasized the positive feedback of these ozone variations on lower stratosphere temperatures, both for the annual cycle[40] and for long-term trends[41]. A further mechanism for ozone variability in the TTL is transport from the extratropics, especially circulations tied to the Asian summer monsoon anticyclone. Some recent calculations point to this in-mixing as a potentially dominant cause of the tropical ozone annual cycle[42–43].

Transport pathways and convective influences

The dominant pathways by which air enters the TTL are via deep convection and three-dimensional transport from circulations that vary both regionally and seasonally. Global models resolve the large-scale tropical circulations[44], with unresolved convective transport effects incorporated through parameterizations. More detailed simulations of transport can be made using three-dimensional trajectory calculations based on analysed meteorological fields, incorporating convective effects by tracking intersections with observed convective systems (based on satellite information[45,46]). Although most convective detrainment occurs near the base of the TTL below ~14 km, extreme overshooting convection can reach the tropopause or even higher[47]. These extreme convective systems are most likely to directly impact stratospheric composition for two reasons. First, large-scale vertical motions are generally upward in the upper TTL above ~15 km, and downward below this level[48]. Hence, detrainment from typical convection at 12–14 km will generally descend back into the middle troposphere, whereas air detrained from extreme convective events will more likely make its way upwards to the stratosphere. Second, very short-lived species affecting ozone concentration (such as bromoform), with a lifetime of several weeks, are more likely to survive the rapid trips to the stratosphere permitted by extreme convective events[49,50]. Trajectory analyses have shown that convection extending high enough to allow transport upwards into the stratosphere occurs primarily over relatively limited geographical regions: the southern continents and the western Pacific in boreal winter, and the Indian subcontinent in the summer[46].

There has been persistent speculation that intense deep convection overshooting the cold point may affect stratospheric humidity directly, thereby decoupling stratospheric water vapour concentration from the tropical tropopause temperature. Modelling studies[51] suggest that plentiful small ice crystals in convective updrafts are likely to prevent detrainment of very dry air from convective overshoots, as hypothesized by Sherwood and Dessler[52]. Moreover, no observational evidence of stratospheric dehydration by overshooting convection has been identified[53]. The *in situ* measurements do provide anecdotal evidence of convective hydration well into the lower stratosphere[2,53,54], and the observed enrichment of water vapour isotopes in the lower stratosphere[55,56] is consistent with detrainment of ice from overshooting convection[57].

Quantification of the global impact of isolated convective hydration events on TTL and stratospheric humidity has proved to be challenging. Cloud-resolving model simulations of particular convective systems suggest the possibility of significant localized moistening of the stratosphere[58], but it is not clear how to extrapolate these results to the global scale. Trajectory analyses including convective injection generally suggest that this process is, at most, of second order importance in comparison with dehydration of air crossing the cold-point tropopause[45,59,60], although perhaps this mechanism is more important in isolated regions, such as near extreme continental convection[3]. Pommereau *et al.*[61] argued that the contrast between lower stratospheric composition over land and oceanic regions suggests significant global impact of overshooting convective systems.

Quantifying the depth of convective penetration and representing extreme convective cloud-top heights in transport calculations represents a significant uncertainty. Geostationary satellite infrared brightness temperature data provides the necessary global coverage and 3-hourly time resolution, but the cloud-top heights derived from this data set have a low bias of 1–2 km (ref. 62). None of the available satellite products provides a complete picture of convective cloud-top heights[63]. NASA's Tropical Rainfall Measuring Mission (TRMM) precipitation radar is biased towards

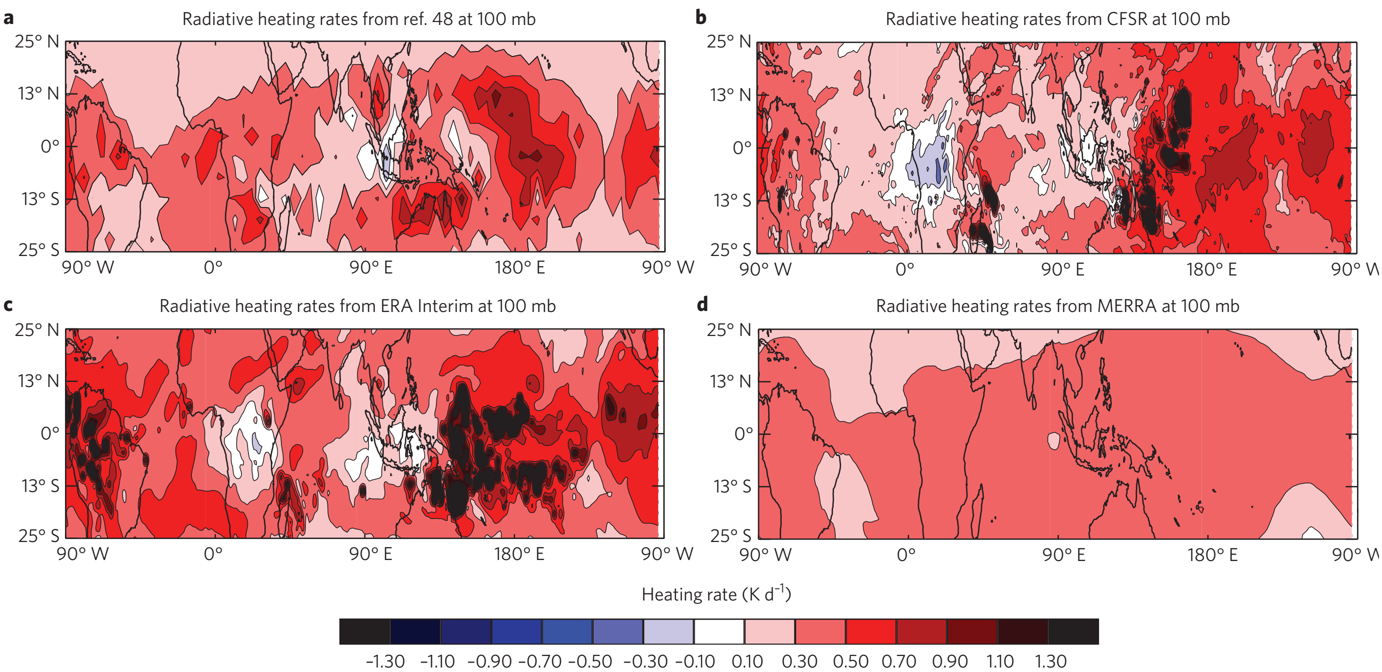

Figure 3 | Large differences exist in current estimates of radiative heating rates in the TTL, and these result in substantial differences in upward transport calculations. These plots compare different estimates of 100 hPa boreal winter climatological heating rates (K day^{-1}). **a**, Results derived from off-line calculations using analysed temperatures, chemical constituents and cloud observations[48]. **b–d**, Analyses from three current-generation meteorological reanalysis systems (ERA-Interim, CSFR and MERRA).

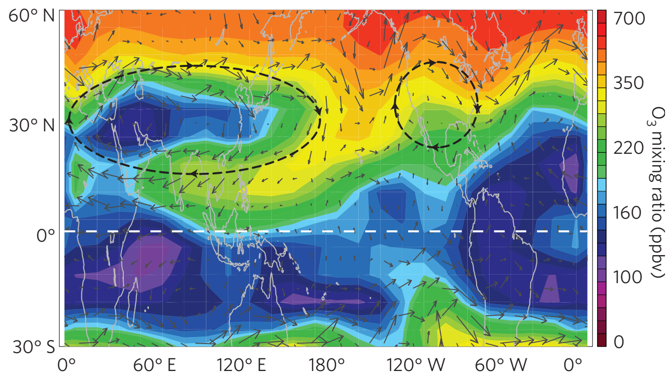

Figure 4 | The horizontal structure of ozone in the lower stratosphere during boreal summer shows the influence of monsoonal circulations on the TTL. Arrows denote the anticyclonic circulations associated with the Asian and North American monsoons; note the ozone transport into the tropics on the eastern flank of the Asian monsoon. The satellite measurements for the 390 K potential temperature level are from the Aura MLS instrument on 4 July 2007.

continental systems with strong updrafts that can loft large graupel. CloudSat, CALIPSO and other satellites in the A-TRAIN only sample at local times around 01:30 and 13:30, and therefore miss the strong late afternoon and early evening peak in continental convective activity. The challenge of modelling convection is further exacerbated by the need for knowledge of the full vertical profiles of mass entrainment and detrainment, which are poorly constrained from observations. Entrainment throughout the depth of the cloud system is especially important for maritime convection with relatively small convective updraft cores[64].

The concentration of water vapour in air entering the tropical stratosphere depends on the pathways taken by air parcels as they transit through the TTL. The slow ascent through the upper TTL occurs along with rapid horizontal transport that carries air parcels large distances through a varying temperature field. Recent trajectory modelling studies have used the 'Lagrangian cold point' paradigm; that is, the water vapour is assumed to be set by the minimum temperature (or, more precisely, the minimum saturation mixing ratio) encountered on an air parcel's journey upwards through the TTL[65,66]. Results from these calculations are sensitive to small differences in the wind and temperature fields employed, and assumptions regarding sub-grid-scale variability[67].

A further uncertainty in TTL transport calculations is the poorly constrained vertical wind speed. Most of the studies use vertical wind speeds calculated in operational meteorological analyses. A kinematic approach, in which the vertical wind is calculated from the horizontal winds using the continuity equation, and a diabatic approach, in which the vertical wind is calculated from the diabatic (mostly radiative) heating rate, have both been used. Comparisons suggest that the latter approach produces a more realistic and less dispersive vertical wind field[68]. An alternative method is to use radiative heating rates calculated from observed temperature, trace constituents and cloud fields[48]; TTL heating rates are sensitive to clouds both locally and at lower altitudes[69,70]. Heating rate calculations from current-generation meteorological analysis systems show large differences (Fig. 3), especially for regional scales, and these in turn produce very different transport estimates[71]. Given the challenge of representing clouds in global models, the uncertainties in heating rates in the meteorological analyses must be considered substantial.

There is new appreciation for the relative importance of quasi-horizontal transport into the TTL from the extratropics, so-called in-mixing. Key components of the circulations at the subtropical edges of the TTL are the summer monsoons, forced as a dynamic response to latent heating in chronic convective regions. These flows consist of large-scale persistent anticyclonic circulations in the subtropics, with the largest and strongest anticyclone tied to the Asian summer monsoon. The monsoon circulations enhance transport into and out of the tropics along the eastern and western flanks (Fig. 4), although the circulation also tends to isolate the air inside the anticyclone.

Because this isolated air inside the Asian anticyclone is often tied to the outflow of deep convection, it is chemically distinct and characteristic of tropospheric composition; for example, high water vapour, low ozone and enhanced pollution[72,73]. Satellite observations[74] furthermore suggest the presence of a persistent aerosol layer of unknown composition near the cold monsoonal tropopause. The three-dimensional monsoon circulation also enhances vertical transport from the troposphere to the stratosphere during summer, as evidenced by trace constituent measurements from satellites[75,76], although the relative contributions of deep convection versus large-scale circulations to this behaviour are poorly constrained.

Clouds and supersaturation in the TTL

The water vapour trajectory calculations described above generally assume that cloud processes immediately remove all vapour in excess of ice saturation from the TTL. However, theoretical work, laboratory experiments and field measurements all suggest that this is an oversimplification[77–79]. It is well established that substantial supersaturation with respect to ice is required to nucleate ice crystals at low temperatures. Homogeneous freezing of aqueous sulphate aerosols, assumed to be the most common type of aerosols in the upper troposphere, does not occur until the relative humidity exceeds ice saturation by about 60–70% at tropical tropopause temperatures[77]. Even heterogeneous ice nucleation on insoluble aerosols typically requires supersaturations of about 10–40%[80]. Further, irreversible dehydration requires sufficiently numerous ice crystals such that they deplete vapour in excess of saturation, as well as large enough ice crystals such that they sediment out of the original cloud layer. Both the vapour depletion and ice sedimentation must occur within the lifetime of the cloud, which will generally be determined by the duration of anomalously cold events driven by synoptic- and meso-scale waves. Measurements of TTL cirrus indicate relatively low ice concentrations (nearly always below 100 L^{-1}[81]) that will not rapidly deplete vapour in excess of saturation.

Estimates of ice supersaturation under the extremely dry TTL conditions have been plagued by persistent discrepancies between water vapour measurements from balloons, aircraft and satellites[82,83]. However, even instruments such as the balloon-borne frostpoint hygrometers, which tend to indicate lower humidities than most of the airborne instruments, suggest that supersaturations up to about 60% occur commonly in the TTL (ref. 84). In fact supersaturation tends to exist even within optically thin cirrus formed *in situ* in the TTL (ref. 85), consistent with the low ice concentrations measured in the clouds. Recent measurements[86] indicate that TTL cirrus ice concentrations are typically not large enough to deplete water vapour above saturation; complete removal of vapour in excess of saturation seems to occur rarely in narrow cloud layers with high ice concentrations produced by homogeneous freezing. Together, these results suggest that air may often ascend across the tropical cold-point tropopause with a considerably higher water vapour concentration than the minimum saturation mixing ratio.

TTL in a changing climate

Several aspects of the TTL are anticipated to evolve in a future changing climate, and indeed some changes have been observed over the past 30 years, the period of available observations during

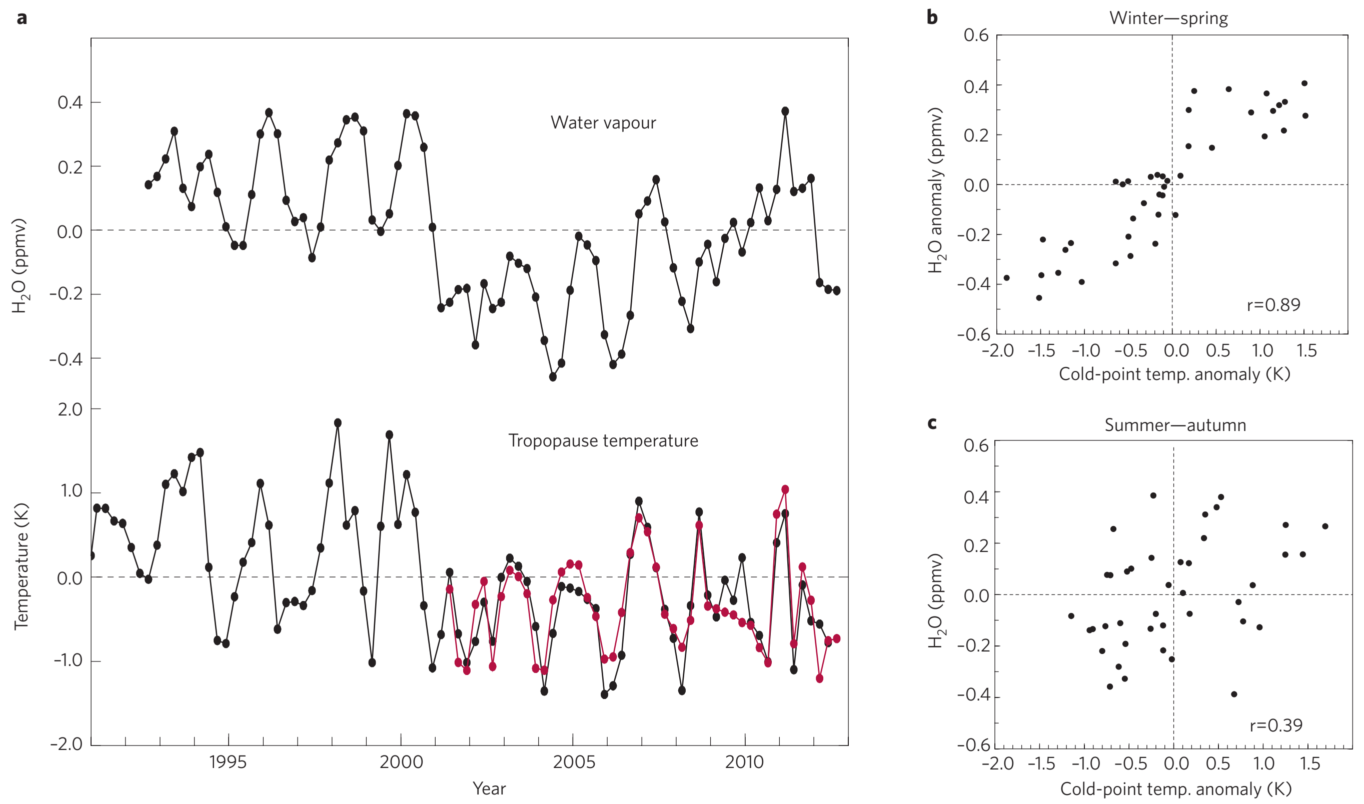

Figure 5 | Interannual changes in global stratospheric water vapour are closely linked to tropical tropopause temperatures. a, Observed interannual changes in lower stratospheric water vapour from satellite measurements over the period 1992–2012 compared with interannual variations in tropical cold-point tropopause temperatures. Water vapour data are de-seasonalized near-global averages at 83 hPa from combined HALOE and MLS satellite measurements. Each dot represents a three-month average. Temperatures are de-seasonalized anomalies derived from radiosonde data (black line) and GPS radio occultation data (red line, for 2001–2012). Results are updated from ref. 93. The approximate two-year variations in the water vapour and temperature anomalies are related to influence of the stratospheric quasi-biennial oscillation (QBO). The correlation between water vapour and temperature for the entire time series is r=0.75. **b**,**c**, Scatter plots of water vapour versus cold-point temperature anomalies (with water vapour lagged by one season), showing high correlation for boreal winter–spring seasons (**b**) and weaker correlation for summer–autumn (**c**).

the satellite measurement era. One important TTL change predicted in chemistry–climate models is an increase in the mean tropical upwelling in the lower stratosphere, as part of the Brewer-Dobson circulation, at a rate of approximately 2% per decade. This appears as a robust result in almost all current models[87], as a consequence of systematic strengthening of the subtropical jets in the lower stratosphere as a response to tropical tropospheric warming, and resulting changes in the dissipation of waves that force tropical upwelling[30,31]. This increased tropical upwelling (and commensurate downwelling in extratropics) has several consequences, including stronger troposphere-to-stratosphere transport and shorter stratospheric lifetimes for chemical tracers.

One effect of stronger upward TTL transport is a decrease in ozone in the tropical lower stratosphere. Although direct observations of upwelling are unavailable and long-term changes cannot be directly measured, observations of tropical lower stratospheric ozone from satellites and balloons show long-term decreases of about 3% per decade from 1984–2009, consistent with increased upwelling predicted by models for this time period[88]. These ozone decreases in turn are likely to result in cooling of the tropical lower stratosphere, and provide an important feedback for temperature and circulation. Recent evidence of seasonal variations in tropical lower stratosphere temperature trends, with the largest cooling during boreal autumn[89], is consistent with similar behaviour in observed ozone changes[41].

Stratospheric water vapour is strongly coupled to tropical cold-point tropopause temperatures on subseasonal to annual time scales, but interannual changes are less well understood. Long-term near-global observations of stratospheric water vapour are available from satellite measurements taken since 1992; longer records are available from isolated balloon measurements at specific locations[90]. The satellite record of lower stratospheric water vapour anomalies (Fig. 5a) shows year-to-year changes of ±0.5 ppmv (~10–15% of background amounts), but small net changes over the 20-year period. The observed water vapour changes are strongly coupled to tropical cold-point temperatures during the cold boreal winter–spring seasons (Fig. 5b), but less so during the summer (Fig. 5c). The water vapour variations observed by satellite from 1992–2012 are not reproduced in chemistry–climate model simulations forced by observed changes in sea surface temperatures, greenhouse gases and ozone-depleting substances[91], suggesting an important component of internal variability within the TTL on interannual time scales that is not captured in current models.

Longer-term historical variations of TTL temperatures, including the cold point, are poorly known because of data uncertainties. Although tropical radiosonde (balloon) measurements began in the 1960s, the historical data record is plagued by sparse sampling, instrumentation changes and data inhomogeneities. Systematic attempts to homogenize and correct the data records have produced some improvements, although typically not with the high vertical resolution needed to constrain detailed TTL structure. For example, the cold point is not analysed in homogenized data sets. Although most chemistry–climate models predict rising and

Box 1 | Processes and trends in current-generation models

Numerical models that explicitly resolve global circulation and chemistry processes (chemistry–climate models, or CCMs) are widely used to simulate the effects of natural and anthropogenic changes on climate. Accurately representing TTL processes in CCMs is critical for simulating present and future climate and variability. Current-generation models (with horizontal resolution of ~100 km and vertical resolution of ~1 km) are able to resolve some key features of the TTL, including the mean thermodynamic structure and circulation within the tropics, the annual cycle in the Brewer-Dobson circulation and large-scale tropical–extratropical coupling[34]. However, several important aspects of the TTL are less well simulated and highly variable among different models, including cold-point temperature (poorly resolved in current models; Fig. B1) and stratospheric water vapour, cloud fractions and ozone throughout the TTL. The influence of overshooting convection is not included in deep convective parameterizations, and the effects of unresolved gravity waves on temperatures, which are important for clouds and dehydration processes, are not simulated. Parameterized representations of cloud processes, which are based on measurements at lower altitudes and higher temperatures, are not necessarily appropriate for cold TTL cirrus.

Interannual changes in the TTL occur because of natural and anthropogenic forcing of the climate system. Two important components of natural variability are: El Niño/Southern Oscillation, a coupled ocean–atmosphere phenomenon with a 2–5-year timescale that extends throughout the troposphere into the lower stratosphere; and the stratospheric quasi-biennial oscillation, a primarily stratospheric phenomenon involving reversal of the tropical zonal winds with an irregular period (~28 months), which extends downwards to the tropical tropopause. These components are simulated (or explicitly forced) with some degree of fidelity in current models, although an explicit goal should be realistic simulations of the relevant processes in free-running coupled models. Model predictions of longer-term TTL evolution associated with greenhouse gas increases include warmer temperatures and increased water vapour in the upper troposphere, higher levels of deep convective outflow, rising and warming of the tropical tropopause, and increased upwelling in the Brewer-Dobson circulation[87,91]. However, these predicted changes must be considered highly uncertain given the limitations of TTL process representation in the models. There are substantial differences among models for other predicted changes, such as for TTL cloud fractions and ozone below the lowermost stratosphere.

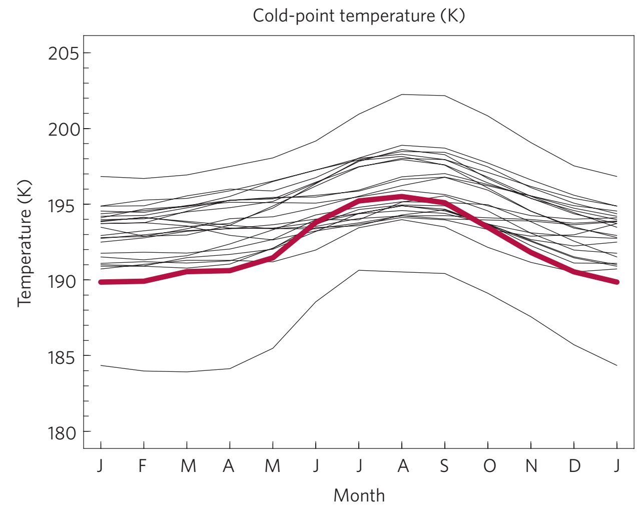

Figure B1 | Seasonal cycle of cold-point tropical tropopause temperature. Black lines show results from the current generation of coupled ocean–atmosphere climate models (specifically, the Coupled Model Intercomparison Project Phase 5 (CMIP5) simulations). The red line shows the cold-point temperature derived from GPS radio occultation data.

warming of the cold-point tropopause and increasing stratospheric water vapour in response to climate change, current observational estimates of past cold-point temperature trends show either cooling or results that are not statistically different from zero[92].

Progress and outlook

Transport within the TTL and coupling with the extratropics is better quantified, based on chemical observations combined with results from trajectory and chemical transport models. The weight of evidence suggests that air in the TTL is often supersaturated, and that the assumption of complete removal of vapour in excess of saturation — often made in global models — should be modified. Multi-year records of constituent and cloud observations from satellites have provided novel perspectives on TTL circulation, and allowed calculations such as comprehensive radiative heating rates. Through this work, the Asian monsoon anticyclone has been identified as the dominant circulation feature during boreal summer, with far-reaching effects on the TTL and global stratosphere.

The most important current research focuses on improving understanding of processes that are poorly resolved in observations and models. The space–time variability of deep convection and its influence on the TTL needs to be better quantified from *in situ* and satellite observations. The transport effects of deep convection, including isolated and extreme overshooting convection, need to be understood and simulated more realistically in models. Quantifying TTL cirrus properties and their dependence on geographic location, temperature and aerosol composition will require further measurements and synthesis.

Large differences in results from meteorological reanalysis systems suggest that models and data assimilation tools for the TTL need improvement. These will be achieved through a better understanding of the relevant processes, on the basis of both global satellite measurements and high-resolution *in situ* measurements. Ensuring the availability of long-term climate-quality data for the TTL, including temperatures, circulation, composition and cloud behaviour, is another priority. These measurements are essential to understand variability and long-term changes in the TTL, and to constrain past and future global model simulations.

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Competing financial interests

The authors declare no competing financial interests.