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Impact of land convection on the thermal structure of the lower stratosphere as inferred from COSMIC GPS radio occultations

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Abstract

Following recent studies evidencing the effect of deep overshooting convection on the chemical composition of the tropical lower stratosphere by injection of tropospheric air across the cold-point tropopause we explore its impact on the thermal structure of the tropical tropopause layer (TTL) and the lower stratosphere using the high-resolution COSMIC GPS radio-occultation temperature measurements spanning from 2006 through 2011. The temperature of the lower tropical stratosphere is shown to display a systematic mean cooling of 0.6 K up to 20 km in the late afternoon in the summer over land compared to oceanic areas where little or no diurnal variation is observed. The temperature cycle is fully consistent with the diurnal cycle and geographical location of deep convective systems reported by the Tropical Rainfall Measurement Mission (TRMM) precipitation radar suggesting strong injection of adiabatically cooled air into the lower tropical stratosphere in the afternoon over tropical continents. But most unexpected is the difference between the southern and Northern Hemispheres, the first displaying systematic larger cooling suggesting more intense convection in the southern than in the northern tropics.

1 Introduction

The tropical tropopause layer (TTL) is recognized for long as the stratospheric “kitchen” playing a major role in stratospheric water vapor and trace gas species controlling stratospheric ozone chemistry and climate. However, the potential impact of the species entering the stratosphere (ice crystals, chemically active very short-lived species or aerosols of different nature) and the dehydration process at the tropopause are highly dependent on the still unclear timescale at which troposphere-to-stratosphere transport (TST) takes place. Indeed, still debated is the contribution of fast overshooting convective updrafts at the global scale in the TTL on a timescale of hours, compared to the slow ascent by radiative heating of the layer above the level

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of neutral buoyancy (LNB) on a timescale of months (Sherwood and Dessler, 2000; Gettelman et al., 2002; Corti et al., 2005; Fueglistaler et al., 2009). The occurrence of convective overshoots reaching 20 km with local upward velocities of up to 50–60 $\mathrm{m\,s^{-1}}$ is known for long (e.g. Vonnegut and Moore, 1958; Burnham, 1970; Roach and James 1972; Cornford and Spavins, 1973; Bowell et al., 1986; Fujita, 1992; Danielsen, 1993). Whereas the existence of deep overshooting above maritime-continental systems as reported by Danielsen (1993) over Northern Australia is generally accepted, several studies (e.g. Highwood and Hoskins 1998; Folkins et al., 1999) suggest that the process is rather episodic and thus of little importance at the global scale. A major attention in the context of TST studies has been paid to the region of the warm pool in the West Pacific, characterized by large-scale slow ascent and minimum Outgoing Longwave Radiation (OLR) used as a proxy of deep convection. However as demonstrated by Alcala and Dessler (2002) from the observations of the precipitation radar (PR) of the Tropical Rainfall Measurement Mission (TRMM), OLR is not a good indicator of cloud penetration into the stratosphere. Indeed, in contrast, the TRMM PR is indicating higher and more frequent "overshooting precipitation features" (OPF) above the tropopause over Africa, South America and Indonesian islands than above oceans (Liu and Zipser, 2005, hereinafter LZ05). The location of the maximum OPFs population is shown to coincide with the areas of maximum flashes frequency reported by the Lightning Imaging Sensor (LIS) also onboard TRMM, indicative of faster updraft velocity over land (Zipser et al., 2006). Furthermore, both OPFs and lightnings are displaying a marked diurnal maximum in the late afternoon over land as opposed to oceanic convection showing very little diurnal change (LZ05). In addition, there is a number of recent observational evidences from the HIBISCUS, TROCCINOX, and SCOUT-O3 European projects in South America, Australia and Africa of frequent deep penetration of tropospheric air and ice crystals up to 19–20 km over land convective systems developing in the afternoon (Nielsen et al., 2007; Corti et al., 2008; Khaykin et al., 2009; Cairo et al., 2010; Pommereau et al., 2011). Such convective updrafts of adiabatically cooled air and ice crystals across the tropopause are now well captured by meso-scale cloud resolving models (Chaboureau et al., 2007; Jensen et al., 2007; Grosvenor et al., 2007; Chemel et al., 2009; Liu et al., 2009) but, because of their non-hydrostatic nature, missed by global meteorological and climate models. While one might argue that overshooting detrainment is episodic and may play a role at local scale only, there is more and more evidences from satellite observations that deep convection above a level of 380 K is not rare phenomena over tropical land and the warm pool ($7 \times 10^6 \mathrm{yr}^{-1}$ between 20° N and 20° S from CALIPSO lidar observations, according to Iwasaki et al. (2010) and might thus have a significant impact on the chemical composition and aerosol loading of the lower stratosphere at the global scale.

The relative importance of the contribution of such continental convective updrafts compared to oceanic areas, with the exception of cyclones displaying also fast uplifts (Tuck et al., 2003), is suggested by the higher concentration of tropospheric trace gas species in the TTL above continents reported by ODIN, MLS, HALOE and MOPITT N_2O, CH_4 and CO profiles measurements (Ricaud et al., 2007, 2009). Another indication of the importance of these is provided by the fast cleansing of stratospheric aerosols up to 20–21 km altitude resulting from the injection of clean tropospheric air during the Southern Hemisphere convective season seen by the CALIPSO lidar (Vernier et al., 2009, 2011).

The question addressed by the present study is how deep convective overshooting could affect the thermal structure of the TTL and the LS. There are already several observational evidences of fast cooling in the afternoon up to 18–19 km in the lower stratosphere over land convective systems, with values ranging from 6–10 K near Borneo in Indonesia (Johnson and Kriete, 1982), 10 K over a "continental-maritime" cloud in Northern Australia (Danielsen, 1993), 4.5–7.5 K in South Eastern Brazil (Pommereau and Held, 2007; Pommereau et al., 2011) and of a mean 2 K cooling at 80 hPa during the two-month monsoon season in West Africa (Cairo et al., 2010). In contrast, with the exception of cyclones where record low cold point temperatures between 183 K and 173 K have been reported at their top (Ebert and Holland, 1991; Danielsen, 1993), the cooling is found generally less intense and at lower altitude over oceans, of the

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order of 2–3 K at 100 hPa from the Atmospheric Infrared Sounder (AIRS) aboard the AQUA satellite (Kim and Dessler, 2004), and of 0.5 K at the cold point from radiosondes in the West Pacific (Sherwood et al., 2003). Using a dry baroclinic model, Highwood and Hoskins (1998) have shown that the thermal properties of the tropopause are indeed related to large-scale convective heating in the troposphere. But the relative contributions of cloud-top radiative cooling, adiabatic lofting, turbulent mixing or wave propagation induced by diabatic heating in the troposphere are still unclear (Randel et al., 2003).

As long as diurnal temperature cycles in the stratosphere are concerned, one should account for the solar tides. Atmospheric tides are global-scale waves of periods that are harmonics of the solar day (Chapman and Lindzen, 1970). The tides are classified into migrating and non-migrating classes. Both tidal classes are generated in the tropical troposphere and propagate as internal gravity waves transporting energy up and away from their source region. These vertically propagating oscillations increase in amplitude with altitude as the atmospheric density decreases. Several studies are showing that migrating tides (components that propagate westward with the apparent motion of the sun) are primarily excited by the zonally symmetric absorption of solar radiation by tropospheric moisture and stratospheric ozone. The non-migrating tides (non sun-synchronous modes which can propagate westward or eastward or stay stationary) are generated by zonally-asymmetric thermal forcing such as planetary boundary layer heat flux or latent heat release in the tropical troposphere, and non-linear interactions between tides and other propagating waves (Teitelbaum and Vial, 1991; Williams and Avery, 1996; Hagan and Forbes, 2002; Lieberman et al., 2004). The most prominent atmospheric tidal mode is the migrating tide, of ~ 1 K amplitude at 30 km at low latitudes (Zeng et al., 2008). The amplitudes of non-migrating tides are significantly smaller in the upper troposphere and lower stratosphere (UTLS), making their observation more difficult. Migrating tides are mainly upper atmospheric features, where for example the amplitude of the diurnal change of temperature exceeds 20 K at 115 km (Zhang et al., 2006). They are well understood and comprehensively documented in

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many observational and theoretical studies and well represented in global ECMWF and NCEP analyses (Pirscher et al., 2010). In contrast, the sources of non-migrating tides, particularly in the UTLS region, are poorly documented and their sources and behaviors still not well known (Oberrheide et al., 2002). Indeed, the available ground-based observations do not allow separating migrating and non-migrating tides (Pirscher et al., 2010 and references therein).

A significant progress in tide studies was achieved thanks to the GPS radio-occultation (RO) technique, utilized by the CHAMP (CHAllenging Minisatellite Payload) mission and the further more elaborated COSMIC mission, providing high resolution temperature profiles in the UTLS. Several studies of the temperature diurnal cycle in the UTLS at low- and mid-latitudes associated with migrating tide are available (e.g. Zeng et al., 2008; Xie et al., 2010; Pircher et al., 2010) but noticeably, without distinguishing between continents and oceans. Here we use the 6-yr 3-dimensional temperature data provided by the COSMIC GPS mission since 2006 for investigating the temperature diurnal cycle in the tropical UTLS and its possible connection with deep convection. Using these data, the geographical distribution of the temperature diurnal cycle in the UTLS is characterized and qualitatively compared to the spatial distribution of deep convection provided by the TRMM PR overshooting precipitation features (OPFs). Since several spectral analyses of phase and amplitude of atmospheric tides are already available, they will not be repeated here, and the study will rather focus on the spatio-temporal distribution of the temperature diurnal cycle. The paper is organized as follows. Section 2 gives a description of the COSMIC RO data set and methodology used. The results are described in Sect. 3, followed by a discussion in Sect. 4 and a summary of conclusions in Sect. 5.

2 Data and methodology

The data used in this study are the temperature profiles derived from the six micro-satellites of the FORMOSAT-3/COSMIC mission (Formosa Satellite

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Mission-3/Constellation Observing System for Meteorology, Ionosphere and Climate) (Anthes et al., 2008) launched in April 2006 jointly by NASA and Taiwan's National Space Organization. The occultation system comprises six low-orbiting satellites evenly distributed in space providing a spatial coverage of 1500–2500 occultations per day. GPS RO measurements are known to be intrinsically calibrated and long-term stable (Steiner et al., 2009). They feature best quality in the upper troposphere and the lower to middle stratosphere (~ 5 km to 35 km altitude); high accuracy and precision (< 1 K for temperature) as well as high vertical resolution from 0.2 km in the troposphere to 1.4 km in the stratosphere (Kursinski et al., 2000; Anthes et al., 2008). Insensitive to clouds and precipitations, the GPS RO technique provides all-weather observations with global coverage and diurnal sampling capability.

The primary product of the GPS RO technique is the bending angle as a function of tangent height, from which the refractivity profile can be retrieved from the top of the atmosphere down to the troposphere using the so-called onion-peeling approach. The temperature and humidity profiles can then be further retrieved from the refractivity (Rocken et al., 1997). Extensive validation and accuracy of the RO measurements against other observations and models have been carried out by many authors (e.g. Kuo et al., 2004; Schreiner et al., 2007).

The data analyzed here are the COSMIC GPS RO level 2 data products called wetPrf (wet profile) which is an interpolated product of 100 m vertical resolution obtained by the one dimensional variational (1DVar) technique together with ECMWF low resolution analyses data. The data include: latitude and longitude of the tangent point, pressure, temperature, water vapor pressure, refractivity profiles and altitude of the perigee a.s.l. The data in NetCDF format are available from the CDACC (COSMIC Data Analysis and Archive Center) website hosted by the University Corporation for Atmospheric Research (UCAR).

The COSMIC data used in this study span from 2006 through 2011, whereas the analysis was restricted to solstice months, i.e. Austral summer (DJF) and Boreal summer (JJA) at low latitudes (25° S–25° N). The interpolated profiles are indexed with respect to the tangent point latitude and longitude at 17 km altitude, the mean altitude of the tropical cold point tropopause. The diurnal variation of temperature is examined by looking at its anomaly compared to the daily mean profile within the 10–35 km height range.

3 Observational results

The relative contributions of migrating and non migrating tides on the temperature diurnal cycle in the tropics are first examined from the zonal-mean temperature anomaly between 10 and 35 km altitude in the southern and northern tropics during JJA and DJF as well as from the diurnal amplitude latitude-height distribution between 25° S–25° N. The influence of continents on the amplitude of the temperature diurnal cycle is then characterized by plotting the difference between land and oceanic areas. Finally, the impact of convective and non-convective regions on the phase and amplitude of the temperature diurnal cycle in the mid- and lower stratosphere and the differences between the Northern and Southern Hemispheres are explored by Hovmoller plots of the diurnal cycle in the austral and boreal summers.

3.1 Zonal-mean profile of temperature diurnal cycle

Figure 1 shows the altitude distribution of the zonally-averaged diurnal temperature anomalies in the Northern (0–25° N) and Southern (0–25° S) tropical belts in JJA and DJF. Both the NH and SH are showing a downward progression of diurnal temperature anomalies of 24 h period, shifted downward by 3 km in the summer compared to the winter in both hemispheres. The temperature variations of largest amplitude are observed around 30 km of gradually decreasing amplitude below. In addition, both hemispheres are showing a second temperature cycle in the lower stratosphere around 17–20 km but in the summer only and of larger amplitude in the South than in the North. This LS cycle is characterized by a cooling in the late afternoon and a nighttime and early morning warming. A further diurnal cycle is observed in the summer in the upper troposphere below 15 km displaying a minimum between 03:00–07:00 LT and a maximum in the early evening between 18:00–22:00 LT.

The downward phase propagation of the temperature anomalies in the mid-stratosphere is the well-known signature of migrating tides already reported by previous studies exploiting COSMIC data and showing an increase of amplitude with height and a local winter maximum (Pirscher et al., 2010; Xie et al., 2010). The temperature diurnal cycle in the lower stratosphere in the local summer of larger amplitude in SH is out of phase with the upper-tropospheric diurnal cycle, showing a warming in the late afternoon due to heat release during the maximum development of convective intensity and the nighttime radiative cooling of clouds.

Figure 2 displays the altitude/latitude distribution of diurnal temperature amplitude in DJF and JJA within the full 25° N–25° S tropical belt. The area of elevated amplitudes above 26 km is the signature of migrating tides. The vertical structure, the amplitude and the winter maximum of these migrating tides fully agree with those derived from the CHAMP (Zeng et al., 2008) and COSMIC data (Pirscher et al., 2010; Xie et al., 2010), and the simulations of the Canadian Middle Atmosphere Model (McLandress, 2002), where the seasonal variation is attributed to the interference of the latitudinal symmetric and anti-symmetric vertically propagating modes of different vertical wavelengths. The northern summer hemisphere shows an additional enhancement between 23–25 km not seen in the SH summer.

A lower thin layer of 0.8 K amplitude is observed in the lowermost stratosphere within 10° S–2° N between the Cold Point Tropopause (CPT) and 20 km in DJF and of 0.6 K within 2° S–5° N in JJA but within 19–20 km only. Remarkably, this layer is decoupled from the higher levels of elevated amplitudes associated with migrating tides, suggesting a different source. Another noticeable feature is the larger amplitude and wider latitude extension of the diurnal cycle in the NH upper troposphere compared to the SH.

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Overall, the temperature diurnal cycle profiles reveal three layers of elevated amplitudes: (i) a broad layer above 26 km altitude bearing the signature of migrating tides of maximum amplitude in the winter; (ii) a thin layer between 17–20 km during the summer season of larger amplitude in the South, and (iii) a third layer in the upper troposphere in the summer but of larger amplitude in the North. While the cause of the diurnal cycle in the first layer related to migrating tide, and in the third layer related to latent heat release and radiative cooling, are well understood, the origin of the temperature cycle in the lower stratosphere during the convective summer season is less clear. The question is whether this cycle is uniform in longitude or dependent on the land-ocean distribution.

3.2 Difference between temperature diurnal cycle above continents and oceans

An answer to this question is provided by the difference between temperature diurnal variations over land and oceans in the SH and NH summers shown in Fig. 3. In the mid-stratosphere, the signature of the zonally-invariant migrating tides is removed, meaning that these are independent of the geographical distribution of continents in both hemispheres. In DJF in the SH tropical belt (left of Fig. 3) the only altitude region where the difference is significant is the UTLS. The warming of the upper troposphere over land between 13:00–18:00 LT is typical of the convective latent heat release and the cooling between 03:00–10:00 LT is a consequence of the nighttime clouds and water vapor radiative cooling. The afternoon cooling between 15–20 LT of the TTL and the lower stratosphere between 17–21 km, drifting downward during the night, is 0.6 K larger over continents than over oceans. The picture in JJA in the NH is somewhat different, showing an afternoon cooling of smaller amplitude and without downward propagation but with the late night warming occurring several hours later. In addition, NH JJA displays a diurnal cycle in the upper levels between 23–26 km, absent in DJF.

For better understanding the observed land/ocean differences and their possible origin we will further explore the zonal variability of the temperature diurnal cycle.

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3.3 Zonal variation of temperature diurnal cycle in the UTLS

Figure 4 shows a Hovmoller diagram of the diurnal temperature anomaly within a 1.5 km layer above the local CPT in NH and SH summer tropical belts. The geo-diurnal temperature pattern in the SH in DJF on the left of Fig. 4 shows a systematic afternoon cooling followed by a nighttime warming of larger amplitude over central Africa and Amazonia and of lesser amplitude over Northern Australia – Indonesian islands and the West Pacific. Remarkably, the continental cold and warm anomalies extend westward to the adjacent oceanic areas.

In NH JJA a similar cycle but of lesser amplitude is observed over West Africa during the monsoon season and of even smaller amplitude over Northern South America and the Gulf of Mexico. In contrast, the Atlantic and the Pacific are displaying a slight noontime cooling followed by a nighttime warming, and the Asian monsoon region an Eastward propagating semi-diurnal cycle. The geo-diurnal temperature pattern in NH JJA is significantly different from SH DJF and displays three diurnal cycle regimes: (i) a land convective type cycle, similar to that observed above SH continents in DJF, over West Africa during the monsoon season and of even smaller amplitude over Northern South America and the Gulf of Mexico; (ii) an oceanic-type cycle with a nighttime warming and a morning cooling, and (iii) a semi-diurnal cycle with Eastward-propagating warm and cold anomalies above Asia and the bay of Bengal during the monsoon season. The semi-diurnal temperature cycle above Asia is present only within the 17–19 km altitude layer.

Overall, the geo-diurnal patterns indicate that the diurnal cycles of largest amplitude in the lower stratosphere are observed above land areas during the summer convective season and that they are of larger amplitude in the Southern Hemisphere. But still open is the question of the process responsible.

Figure 5 displays the geographical distribution of the temperature difference in a 1.5 km thick layer above the local CPT between the late afternoon (18:00 LT ± 2 h) and the morning (10:00 LT ± 2 h), the periods of maximum and minimum convective

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activity over land (LZ05). The COSMIC profiles are binned into 5° × 5° boxes, each comprising over 300 profiles on average. Also shown in Fig. 5 is the geographical distribution of the frequency of occurrence of TRMM PR OPFs above 14 km (adopted from LZ05). The coincidence between the location of late afternoon cooling and overshooting convection is nearly perfect in DJF. The largest OPFs population and highest amplitudes of lower stratosphere cooling are observed in the south tropical belt over Africa and South America. Although more dispersed and of smaller amplitude, cooling features are also observed over the Indonesian islands and Northern Australia. During JJA the afternoon cooling features are also seen over convective areas of Central and West Africa, Northern South-America and Indonesian Islands, but absent over the convective Central America and India during monsoon season.

In summary, significant afternoon cooling in the lower stratosphere is confined to the deep convection areas but over land only and particularly large near the equator and in the SH tropics.

3.4 Temperature diurnal cycle in the mid-stratosphere

As already seen in Fig. 3 there is a temperature diurnal cycle displaying an afternoon maximum and an early morning minimum within the 23–26 km layer in the NH summer, which is absent in the SH summer. Since the altitude of this feature is too high for being related to convective lofting, another explanation is required. The geodiurnal temperature variations in the 23–26 km layer for DJF in the SH and JJA in the NH are shown in Fig. 6. The SH summer displays late evening maxima and early noon minima over Africa, the East Atlantic and South America implying afternoon warming, but little change in the Pacific and the Indian Ocean, where limited diurnal variation is observed. The geodiurnal pattern is rather different over the northern tropics in JJA where a diurnal variation of 1.1 K peak-to-peak amplitude is observed above the desert regions of Sahara and Saudi Arabia, and though of lesser amplitude, a semi-diurnal cycle over the Asian monsoon region. The diurnal cycle over desert areas is showing a maximum temperature in the early evening and a minimum in the early morning

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implying a warming during the full day. Note that this diurnal cycle of large amplitude and longitude extension above Sahara and Saudi Arabia is strongly weighing the zonal mean land-ocean difference shown in Fig. 3.

The daytime warming over desert regions is attributed to long and short waves absorption by ozone at the altitude of its maximum concentration following the OLR increase from 290 to 370 $\mathrm{W\,m^{-2}}$ during daytime after the solar heating of the surface (Schmetz and Liu, 1988; Smith and Rutan, 2003) and shortwave solar light reflection due to the 0.31 albedo at 630 nm of desert areas (Harrison et al., 1990; Minnis and Young, 2002). The peak-to-peak amplitude of the daytime heating of 1.1 K around 23–26 km above the NH deserts is consistent with model estimates of heating by longwave ozone absorption (Gettelman et al., 2004).

Of similar radiative origin, but of lesser amplitude and shifted in time, is the diurnal cycle over Asia during the monsoon season and over Africa and South America during the convective summer, displaying a minimum temperature around or shortly after noontime followed by a fast heating in the afternoon and a slow nighttime and morning cooling. Such behavior is consistent with the daytime development of land convective systems above which the OLR is decreasing in the morning and the albedo is reaching a value of 1 at the top of storm anvils in the afternoon.

The mid-stratosphere temperature diurnal cycle is fully explained by the radiative heating of the ozone layer by short and long wave radiation. The only remaining question is the process responsible for the temperature diurnal cycle in the lower stratosphere and the TTL above land convective systems.

4 Impact of convection on the thermal structure of the lower stratosphere

The collocation in space and time of the afternoon cooling of the lower stratosphere and the maximum development of land convection indicates that the latter is responsible for the observed temperature diurnal cycle. The well known process is the injection of adiabatically cooled air across the tropopause by overshooting turrets followed by turbulent

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mixing with warmer stratospheric air, as suggested by Danielsen (1982) and further demonstrated by Danielsen (1993), Sherwood and Dessler (2000) and the many other studies quoted in the introduction. In addition, the turbulent mixing may be enhanced by the breaking of Kelvin waves as proposed by Fujiwara et al. (2003) thereby extending vertically the mixing layer. The colder and heavier air masses are then descending, resulting in an adiabatic heating of air above the cold point, as seen in the temperature diurnal variation over land (Fig. 3). Note that the cold anomaly persists until 02:00 LT during the night, which is consistent with Liu and Zipser (2009) showing higher CO concentration in the TTL at 01:30 than at 13:30 due to the ability of land convective systems to maintain their overshooting activity until the early night. The resulting systematic diurnal cycle is consistent with those already reported by Johnson and Kriete (1982) over Borneo or Pommereau et al. (2011) over Brazil and Cairo et al. (2010) over West Africa.

An important consideration is the altitude reached by the overshoots, which defines the top altitude of the influence of the troposphere that is the top of the TTL. Note that though the probability for overshoot to reach a given level decreases exponentially with height (LZ05), the amplitude of the adiabatic cooling increases with altitude. The COSMIC RO temperatures are showing a maximum cooling at 19.2 km, i.e. 2 km higher than the mean CPT, during the summer in the Southern Hemisphere. This altitude is very consistent with the observations of geyser-like injections of ice crystals at about the same level (Kelly et al., 1993; Nielsen et al., 2007; Corti et al., 2008; De Reus et al., 2008; Khaykin et al., 2009). According to COSMIC RO temperatures the cold anomaly over land extends up to 20 km in the SH, which is also very consistent with the zonal mean convective cleansing of the aerosols by injection of clean air in the lower stratosphere up to 20 km during the SH summer reported by Vernier et al. (2011) from CALIPSO lidar observations, suggesting that, at least above continents and at least in the SH, the top TTL might reach 20 km.

Various other processes might also impact the thermal structure of the lower stratosphere, such as cloud-top radiative cooling, Kelvin or Rossby waves, quasi-horizontal

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eddies or gravity waves of short vertical wavelengths. However convective overshoots followed by turbulent mixing is the only process allowing a fast cooling of the lower stratosphere by several degrees on a time scale of a few hours required for explaining the systematic diurnal cycle. In addition, processes other than fast cross-tropopause updrafts cannot explain the already cited injection of ice crystals and cleansing of the aerosols or the increase of tropospheric trace gases concentration in the stratosphere over land (Ricaud et al., 2007; Schoeberl et al., 2006), moreover restricted to the Austral summer (Ricaud et al., 2009), which all require a transfer of mass and not of energy only. Furthermore convective overshooting is a well-understood process fully captured by non-hydrostatic cloud resolving models (Jensen et al., 2007; Chaboureau et al., 2007; Grosvenor et al., 2007; Liu et al., 2011) providing confidence to the above interpretation.

An important finding is the larger amplitude of the temperature diurnal cycle in the lower stratosphere over land convective systems in the southern than in the northern tropics, compatible with the restriction to the austral summer of the cleansing of the aerosols and the higher increase of trace gases concentration in the LS during the austral summer compared to the boreal summer. This would imply much more intense convective overshooting in the Southern Hemisphere, which is consistent with TRMM PR observations indicating that most vigorous convection is occurring first above the Congo basin in Central Africa and second over Amazonia during the austral summer (Zipser et al., 2007; Liu and Zipser, 2009). A plausible origin of this difference is the known lesser abundance of anthropogenic aerosols and desert dust over African and South-American rain forests compared to the Northern Hemisphere tropics which, because of their opposite microphysical and radiative effects, can suppress convection intensity at large concentrations (Koren et al., 2008; Rosenfeld et al., 2008; Tao et al., 2012).

Finally, although overshooting of adiabatically cooled air masses appears a consistent explanation, a contribution of non-migrating tides, that is of internal gravity waves to the temperature diurnal cycle cannot be totally ruled out. Many authors have studied the forcing of non-migrating tides in the upper atmosphere following convective latent heat release (e.g. Forbes et al., 1997; Hagan and Forbes, 2002; Oberheide et al., 2002). However none of them provided clear information on the temperature response in the lower stratosphere since such signal is not expected to be detectable at this altitude. The effect of non-migrating tides might indeed explain some of the observed features such as the layer of enhanced diurnal amplitudes at 19–20 km near the equator in JJA (Fig. 2) in the absence of a signature immediately above the CPT. A similar amplitude enhancement is also observed in DJF at the same elevation (occurring above the sharp layer in the TTL), thus likely of the same origin. A typical property of non-migrating tides is reflected by the zonal propagation of temperature anomalies seen in the UTLS geodiurnal pattern (Fig. 4), showing the extension of the land diurnal cycle over the adjacent oceans. A very similar phenomenon was described by Yang and Slingo (2001), who detected on the basis of multi-satellite data sets of brightness temperature a spread of land convection diurnal cycle several hundreds kilometers over the adjacent oceans, an effect tentatively attributed to the gravity waves of varying depth.

5 Summary

The six-year COSMIC RO temperature profiles have been used to characterize the temperature diurnal variation in the tropical UTLS and investigate its relationship with continental convection. The analysis reveals the signature of the well known migrating solar tides in the upper stratosphere above 26 km altitude and non-migrating signals at several levels below: between 23–26 km in the mid-stratosphere, between the cold point tropopause and 20 km in the lower stratosphere and in the upper troposphere. While the upper level migrating signal is shown to be independent of the geographical distribution of continents, all others are found enhanced above land compared to oceans. The mid-stratosphere non-migrating tides signatures are fully explained by the radiative heating of the ozone layer by absorption of short- and long-wave radiation

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from the surface. They are particularly large over Sahara and Saudi Arabia deserts and land areas in phase with the strong diurnal cycle of convective systems but almost absent above oceanic regions where convection does not show diurnal variation.

In the TTL, the temperature diurnal cycles are showing strong afternoon cooling collocated with land convective systems developing in the afternoon during the summer as detected by the TRMM precipitation radar. The largest TTL temperature cycles are observed in the Southern tropical belt above Africa, South America and Indonesian Islands during the austral summer, and of lesser amplitude above West Africa and Northern South America in the Northern tropical belt during the boreal summer. The largest amplitudes in the boreal summer are observed above Asia, displaying a semi-diurnal cycle with a fast morning cooling.

The systematic afternoon cooling between the tropopause and 20 km altitude over land is shown fully compatible with the systematic cross-tropopause injections of adiabatically cooled air by convective overshooting, a non-hydrostatic process suggested by Danielsen (1982), now well captured by mesoscale Cloud Resolving Models. The reality of this hypothesis is strengthened by the many recent observations of ice crystals, tropospheric trace gases and clean air injections in the lower stratosphere during the austral summer, which all require a mass transfer and not of energy only.

The further information delivered by the COSMIC RO temperatures is the systematically larger impact of convection in the Southern tropics compared to the Northern, which, together with the similar conclusions for trace gases and clean air injections, would imply more vigorous overshooting over SH land areas. Such difference between the two hemispheres is tentatively attributed to the larger amount of anthropogenic aerosols and desert dust in the northern tropics compared to African and South American rain forests areas in the south, although the microphysical and radiative processes involved will need further investigation.

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References

Alcala, C. M. and Dessler, A. E.: Observations of deep convection in the tropics using the Tropical Rainfall Measuring Mission (TRMM) precipitation radar, J. Geophys. Res., 107, 4792, doi:10.1029/2002/D002457, 2002.

Anthes, R. A., Bernhardt, P. A., Chen, Y., Cucurull, L., Dymond, K. F., Ector, D., Healy, S. B., Ho, S.-P., Hunt, D. C., Kuo, Y.-H., Liu, H., Manning, K., McCormick, C., Meehan, T. K., Randel, W. J., Rocken, C., Schreiner, W. S., Sokolovskiy, S. V., Syndergaard, S., Thompson, D. C., Trenberth, K. E., Wee, T.-K., Yen, N. L., and Zeng, Z.: The COSMIC/FORMOSAT-3 mission early results, B. Am. Meteorol. Soc., 89, 313–333, doi:10.1175/BAMS-89-3-313, 2008.

Browell, E. V., Danielsen, E. F., Holton, J. R., Hoskins, B. J., Johnson, D. R., Kley, D., Krieger A. J., Megie, G., Newell, R. E., and Vaughan, G.: Strat-Trop exchange, Atmospheric ozone, 1985, 151–240, WMO Global Ozone Research and Monitoring Project, Report no. 16, 1986.

Burnham J.: Atmospheric gusts – a review of the results of some recent research at the royal aircraft establishment, Mon. Weather. Rev., 98, 723–734, 1970.

Cairo, F., Pommereau, J. P., Law, K. S., Schlager, H., Garnier, A., Fierli, F., Ern, M., Streibel, M., Arabas, S., Borrmann, S., Berthelier, J. J., Blom, C., Christensen, T., D'Amato, F., Di Donfrancesco, G., Deshler, T., Diedhiou, A., Durry, G., Engelsen, O., Goutail, F., Harris, N. R. P.,

ACPD

13, 1–31, 2013

Impact of land convection

S. M. Khaykin et al.

Kerstel, E. R. T., Khaykin, S., Konopka, P., Kylling, A., Larsen, N., Lebel, T., Liu, X., MacKenzie, A. R., Nielsen, J., Oulanowski, A., Parker, D. J., Pelon, J., Polcher, J., Pyle, J. A., Ravegnani, F., Rivière, E. D., Robinson, A. D., Röckmann, T., Schiller, C., Simões, F., Stefanutti, L., Stroh, F., Some, L., Siegmund, P., Sitnikov, N., Vernier, J. P., Volk, C. M., Voigt, C., von Hobe, M., Viciani, S., and Yushkov, V.: An introduction to the SCOUT-AMMA stratospheric aircraft, balloons and sondes campaign in West Africa, August 2006: rationale and roadmap, Atmos. Chem. Phys., 10, 2237–2256, doi:10.5194/acp-10-2237-2010, 2010.

Chaboureau, J.-P., Cammas, J.-P., Duron, J., Mascart, P. J., Sitnikov, N. M., and Voessing, H.-J.: A numerical study of tropical cross-tropopause transport by convective overshoots, Atmos. Chem. Phys., 7, 1731–1740, doi:10.5194/acp-7-1731-2007, 2007.

Chapman, S. and Lindzen, R. S.: Atmospheric Tides Thermal and Gravitational, D. Reidel, Dordrecht, Holland, 200 pp., 1970.

Chemel, C., Russo, M. R., Pyle, J. A., Sokhi, R. S., and Schiller, C.: Quantifying the imprint of a severe hector thunderstorm during ACTIVE/SCOUT-O3 onto the water content in the upper troposphere/lower stratosphere, Mon. Weather Rev., 137, 2493–2514, doi:10.1175/2008MWR2666.1, 2009.

Cornford S. G. and Spavins, C. S.: Some measurements of cumulonimbus tops in the premonsoon season in North-East India, Meteorol. Mag., 102, 314–332, 1973.

Corti, T., Luo, B. P., and Peter, T.: Mean radiative energy balance and vertical mass fluxes in the equatorial upper troposphere and lower stratosphere, Geophys. Res. Lett., 32, L06802, doi:10.1029/2004GL021889, 2005.

Corti, T., Luo, B. P., Fu, Q., Vömel, H., and Peter, T.: The impact of cirrus clouds on tropical troposphere-to-stratosphere transport, Atmos. Chem. Phys., 6, 2539–2547, doi:10.5194/acp-6-2539-2006, 2006.

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., Vossing, H. J., Borrmann, S., and Peter, T.: Unprecedented evidence for deep convection hydrating the tropical stratosphere, Geophys. Res. Lett., 35, L10810, doi:10.1029/2008GL033641, 2008.

Danielsen, E. F.: A dehydration mechanism for the stratosphere, Geophys. Res. Lett., 9, 605–608, 1982.

CC BY

Danielsen E. F.: In situ evidence of rapid, vertical, irreversible transport of lower tropospheric air into the lower stratosphere by convective cloud turrets and by large scale up welling in tropical cyclones, J. Geophys. Res, 98, 8665–8681, 1993.

de Reus, M., Borrmann, S., Bansemer, A., Heymsfield, A. J., Weigel, R., Schiller, C., Mitev, V., Frey, W., Kunkel, D., Kürten, A., Curtius, J., Sitnikov, N. M., Ulanovsky, A., and Ravegnani, F.: Evidence for ice particles in the tropical stratosphere from in-situ measurements, Atmos. Chem. Phys., 9, 6775–6792, doi:10.5194/acp-9-6775-2009, 2009.

Ebert, E. E. and Holland, G. J: Observations of record cold cloud-top temperature in tropical cyclone Hilda (1990), Mon. Weather. Rev., 120, 2240–2251, 1992.

Folkins, I., Loewenstein, M., Podolske, J., Oltmans, S. J., and Profitt, M.: A barrier to vertical mixing at 14 km in the tropics: evidence from ozonesondes and aircraft measurements, J. Geophys. Res., 104, 22,095–22,102, 1999.

Forbes, J. M., Hagan, M. E., Zhang, X., and Hamilton, K.: Upper atmosphere tidal oscillations due to latent heat release in the tropical troposphere, Ann. Geophys., 15, 1165–1175, doi:10.1007/s00585-997-1165-0, 1997.

Fueglistaler, S., Dessler, A. E., Dunkerton, T. J., Folkins, I., Fu, Q., and Mote, P. W.: The tropical tropopause layer, Rev. Geophys., 47, RG1004, doi:10.1029/2008RG000267, 2009.

Fujita, T. T.: Mystery of Severe Storms, during the 50 years 1942–1992, Wind Research Laboratory, Department of Geophysical Sciences, University of Chicago, 1992.

Fujiwara, M., Yamamoto, M. K., Hashiguchi, H., Horinouchi, T., and Fukao, S.: Turbulence at the tropopause due to breaking Kelvin waves observed by the Equatorial Atmosphere Radar, Geophys. Res. Lett., 30, 1171, doi:10.1029/2002GL016278, 2003.

Gage, K. S., McAfee, J. R., Carter, D. A., Ecklund, W. L., Riddle, A. C., Reid, A. C., and Balsley, B. B.: Long-term mean vertical motion over the tropical Pacific: wind-profiling Doppler radar measurements, Science, 254, 1771–1773, 1991.

Gettelman, A., Salby M. L., and Sassi, F: The distribution and influence of convection in the tropical tropopause region, J. Geophys. Res., 107, 4080, doi:10.1029/2001JD001048, 2002.

Gettelman, A., de F. Fujiwara, P. M., Fu, Q., Vömel, H., Gohar, L. K., Johanson, C., and Aaammerman, M.: Radiation balance of the tropical layer, J. Geophys. Res. 109, D07103, doi:10.1029/2003JD004190, 2004.

Grosvenor, D. P., Choularton, T. W., Coe, H., and Held, G.: A study of the effect of overshooting deep convection on the water content of the TTL and lower stratosphere from Cloud

Resolving Model simulations, Atmos. Chem. Phys., 7, 4977–5002, doi:10.5194/acp-7-4977-2007, 2007.
Hagan, M. E. and J. M. Forbes: Migrating and nonmigrating diurnal tides in the middle and upper atmosphere excited by tropospheric latent heat release, J. Geophys. Res., 107, 4754, doi:10.1029/2001JD001236, 2002.
Harrison, E. F., Minnis, P., Barkstrom, B. R., Ramanathan, V., Cess, R. D., and Gibson, G. G.: Seasonal variation of cloud radiative forcing derived from the Earth Radiation Budget Experiment, J. Geophys. Res., 95, 18687–18703, 1990.
Highwood, E. J. and Hoskins, B. J.: The tropical tropopause, Q. J. Roy. Meteor. Soc., 124, 1579–1604, 1998.
Iwasaki, S., Shibata, T., Nakamoto, J., Okamoto, H., Ishimoto, H., and Kubota, H.: Characteristics of deep convection measured by using the A-train constellation, J. Geophys. Res., 115, D06207, doi:10.1029/2009JD013000, 2010.
Jensen, E. J., Ackerman, A. S., and Smith J. A.: Can overshooting convection dehydrate the tropical tropopause layer?, J. Geophys. Res., 112, D11209, doi:10.1029/2006JD007943, 2007.
Johnson, R. H. and Kriete, D. C.: Thermodynamic and circulation characteristics of winter monsoon tropical mesoscale convection, Mon. Weather Rev., 10, 1898–1911, 1982.
Kelly, K. K., Profitt, M. H., Chan, K. R., Loewenstein, M., Podolske, J. R., Strahan, S. E., Wilson, J. C., and Kley, D.: Water vapor and cloud measurements over Darwin during STEP 1987 tropical mission, J. Geohys. Res., 98, 8713–8723, 1993.
Khaykin, S., Pommereau, J.-P., Korshunov, L., Yushkov, V., Nielsen, J., Larsen, N., Christensen, T., Garnier, A., Lukyanov, A., and Williams, E.: Hydration of the lower stratosphere by ice crystal geysers over land convective systems, Atmos. Chem. Phys., 9, 2275–2287, doi:10.5194/acp-9-2275-2009, 2009.
Kim, H. and Dessler, A. E.: Observations of convective cooling in the tropical tropopause layer in AIRS data, Atmos. Chem. Phys. Discuss., 4, 7615–7629, doi:10.5194/acpd-4-7615-2004, 2004.
Kley, D., Schmeltkopf, A. L., Kelly, K., Winkler, R. H., Thompson, T. L., and McFarland, M.: Transport of water vapor through the tropical tropopause, Geophys. Res. Lett., 9, 617–620, 1982.
Koren, I., Martins, J. V., Remer, L. A., and Afargan H.: Smoke invigoration versus inhibition of clouds over the Amazon, Science, 946–949, doi:10.1126/science.1159185, 2008.

cc BY

Kuo, Y. H., Wee, T. K., Sokolovskiy, S., Rocken, C., Schreiner, W., Hunt, D., and Anthes, R. A.: Inversion and error estimation of GPS radio occultation data, J. Meteorol. Soc. Jpn, 82, 507–531, 2004.

Kursinski, E. R., Hajj, G. A., Schofield, J. T., Linfield, R. P., and Hardy, K. R.: Observing Earth's atmosphere with radio occultation measurements using the Global Positioning System. J. Geophys. Res., 102, 23429–23465, 1997.

Lieberman, R. S.: Non migrating diurnal tides in the equatorial middle atmosphere, J. Atmos. Sci., 48, 1112–1123, 2004.

Liu, C. and Zipser, E. J.: Global distribution of convection penetrating the tropical tropopause, J. Geophys. Res., doi:10.1029/2005JD006063, 2005.

Liu, C. and Zipser, E. J.: Implications of the day versus night differences of water vapor, carbon monoxide, and thin cloud observations near the tropical tropopause, J. Geophys. Res., 114, D09303, doi:10.1029/2008JD011524, 2009.

Liu, X. M., Rivière, E. D., Marécal, V., Durry, G., Hamdouni, A., Arteta, J., and Khaykin, S.: Stratospheric water vapour budget and convection overshooting the tropopause: modelling study from SCOUT-AMMA, Atmos. Chem. Phys., 10, 8267–8286, doi:10.5194/acp-10-8267-2010, 2010.

McLandress, C.: The seasonal variation of the propagating diurnal tide in the mesosphere and lower thermosphere, Part II: The role of tidal heating and zonal mean winds, J. Atmos. Sci., 59, 907–922, 2002.

Minnis, P. and Young, D. F.: Surface and clear-sky albedos and emissivities from MODIS and VIRS with application to SEVIRI, Extended abstract in proc. of the 2002 EUMETSAT Meteorological Satellite Conference Dublin, Ireland, 2–6 September, 2002.

Nielsen, J. K., Larsen, N., Cairo, F., Di Donfrancesco, G., Rosen, J. M., Durry, G., Held, G., and Pommereau, J. P.: Solid particles in the tropical lowest stratosphere, Atmos. Chem. Phys., 7, 685–695, doi:10.5194/acp-7-685-2007, 2007.

Oberheide, J., Hagan, M. E., Roble, R. G., and Offermann, D.: Sources of nonmigrating tides in the tropical middle atmosphere, J. Geophys. Res., 107, 4567, doi:10.1029/2002JD002220, 2002.

Pirscher, B., Foelsche, U., Borsche, M., Kirchengast, G., and Kuo, Y.-H.: Analysis of migrating diurnal tides detected in FORMOSAT-3/COSMIC temperature data, J. Geophys. Res., 115, D14108, doi:10.1029/2009JD013008, 2010.

CC BY

Pommereau, J.-P. and Held, G.: Is there a stratospheric fountain?, Atmos. Chem. Phys. Discuss., 7, 8933–8950, doi:10.5194/acpd-7-8933-2007, 2007.

Pommereau, J.-P., Garnier, A., Held, G., Gomes, A. M., Goutail, F., Durry, G., Borchi, F., Hauchecorne, A., Montoux, N., Cocquerez, P., Letrenne, G., Vial, F., Hertzog, A., Legras, B., Pisso, I., Pyle, J. A., Harris, N. R. P., Jones, R. L., Robinson, A. D., Hansford, G., Eden, L., Gardiner, T., Swann, N., Knudsen, B., Larsen, N., Nielsen, J. K., Christensen, T., Cairo, F., Fierli, F., Pirre, M., Marécal, V., Huret, N., Rivière, E. D., Coe, H., Grosvenor, D., Edvarsen, K., Di Donfrancesco, G., Ricaud, P., Berthelier, J.-J., Godefroy, M., Seran, E., Longo, K., and Freitas, S.: An overview of the HIBISCUS campaign, Atmos. Chem. Phys., 11, 2309–2339, doi:10.5194/acp-11-2309-2011, 2011.

Randel, W. J., Wu, F., and Rivera Ríos, W.: Thermal variability of the tropical tropopause region derived from GPS/MET observations, J. Geophys. Res., 108, 4024, doi:10.1029/2002JD002595, 2003.

Ricaud, P., Barret, B., Attié, J.-L., Motte, E., Le Flochmoën, E., Teyssèdre, H., Peuch, V.-H., Livesey, N., Lambert, A., and Pommereau, J.-P.: Impact of land convection on troposphere-stratosphere exchange in the tropics, Atmos. Chem. Phys., 7, 5639–5657, doi:10.5194/acp-7-5639-2007, 2007.

Ricaud, P., Pommereau, J.-P., Attié, J.-L., Le Flochmoën, E., El Amraoui, L., Teyssèdre, H., Peuch, V.-H., Feng, W., and Chipperfield, M. P.: Equatorial transport as diagnosed from nitrous oxide variability, Atmos. Chem. Phys., 9, 8173–8188, doi:10.5194/acp-9-8173-2009, 2009.

Roach, W. T. and James, B. F. : A climatology of the potential vertical extent of giant cumulonimbus in some selected areas, Meteorol. Mag. 101, 161–181, 1972.

Rocken, C., Anthes, R., Exner, M., Hunt, D., Sokolovskiy, S., Ware, R., Gorbunov, M., Schreiner, W., Feng, D., Herman, B., Kuo, Y.-H., and Zou, X.: Analysis and validation of GPS/MET data in the neutral atmosphere, J. Geophys. Res., 102, 29849–29866, 1997.

Rosenfeld, D., Lohmann, U., Raga, G. B., O'Dowd, C. D., Kulmala, M., Fuzzi, S., Reissell, A., and Andreae, M. O.: Flood or drought: how do aerosols affect precipitation?, Science, 321, 1309–1313, 2008.

Schmetz J. and Liu, Q.: Outgoing long wave radiation and its diurnal variation at regional scles derived from Meteosat, J. Geophys. Res., 93; 11192–11204, 1988.

Schreiner, W., Rocken, C., Sokolovskiy, S., Syndergaard, S., and Hunt, D.: Estimates of the precision of GPS radio occultations from the COSMIC/FORMOSAT-3 mission, Geophys. Res. Lett., 34, L04808, doi:10.1029/2006GL027557, 2007.

Sherwood, S. C. and Dessler, A. E.: Convective mixing near the tropical tropopause: insights from seasonal variations, J. Atmos. Sci., 60, 2674–2685, 2003.

Sherwood, S. C., Horinouchi, T., and Zeleznik, H. A.: Convective impact on temperatures observed near the tropical tropopause, J. Atmos. Sci., 60, 1847–1856, 2003.

Schoeberl, M. R., Duncan, B. N., Douglass, A. R., Waters, J., Livesey, N., Read, W., and Filipiak, M.: The carbon monoxide tape recorder, Geophys. Res. Lett., 33, L12811, doi:10.1029/2006GL026178, 2006.

Smith, G. L. and Rutan, D. A.: The diurnal cycle of outgoing longwave radiation from Earth radiation budget experiment measurements, J. Atmos. Sci., 60, 1529–1542, doi:10.1175/2997.1, 2003.

Steiner, A. K., Kirchengast, G., Lackner, B. C., Pirscher, B., Borsche, M., and Foelsche, U.: Atmospheric temperature change detection with GPS radio occultation 1995 to 2008, Geophys. Res. Lett., 36, L18702, doi:10.1029/2009GL039777, 2009.

Tao, W.-K., Chen, J.-P., Li, Z., Wang, C., Zhang, C.: Impact of aerosols on convective clouds and precipitation, Rev. Geophys., 50, RG2001, doi:10.1029/2011RG000369, 2012.

Teitelbaum, H. and Vial, F.: On tidal variability by nonlinear interaction with planetary wave, J. Geophys. Res., 96, 14169–14178, 1991.

Tuck, A. F., Hovde, S. J., Kelly, K. K., Mahoney, M. J., Proffitt, M. H., Richard, E. C., and Thompson, T. L.: Exchange between the upper tropical troposphere and the lower stratosphere studied with aircraft observations, J. Geophys. Res., 108, 4734, doi:10.1029/2003JD003399, 2003.

Vernier, J. P., Pommereau, J. P., Garnier, A., Pelon, J., Larsen, J. Nielsen, N., Christensen, T., Thomasson, L. W., Leblanc, T., and McDermid, I. S.: Troposphere stratosphere transport in the tropics from CALIOP lidar aerosols measurements, J. Geophys. Res., 114, D00H10, doi:10.1029/2009JD011946, 2009.

Vernier, J.-P., Pommereau, J.-P., Thomason, L. W., Pelon, J., Garnier, A., Deshler, T., Jumelet, J., and Nielsen, J. K.: Overshooting of clean tropospheric air in the tropical lower stratosphere as seen by the CALIPSO lidar, Atmos. Chem. Phys., 11, 9683–9696, doi:10.5194/acp-11-9683-2011, 2011.

Vonnegut, B. and Moore, C. B.: Giant Electrical storms, Recent Advances in Atmospheric Electricity, Pergamon Press, London, 399–411, 1958.
Williams, C. R. and Avery, S. K.: Diurnal nonmigrating tidal oscillations forced by deep convetive clouds, J. Geophys. Res., 101, 4079–4091, 1996.
Xie, F., Wu, D. L., Ao, C. O., and Mannucci, A. J.: Atmospheric diurnal variations observed with GPS radio occultation soundings, Atmos. Chem. Phys., 10, 6889–6899, doi:10.5194/acp-10-6889-2010, 2010.
Yang G.-Y. and Slingo J.: The diurnal cycle in the tropics, Mon. Weather Rev., 129, 784–801, 2001.
Zhang, X., Forbes, J. M., Hagan, M. E., Russell III, J. M., Palo, S. E., and Mlynczak, M.: Monthly tidal temperatures 20–120 km from TIMED/SABER, J. Geophys. Res., 111, A10S08, doi:10.1029/2005JA011504, 2006.
Zeng, Z., Randel, W., Sokolovskiy, S., Deser, C., Kuo, Y.-H., Hagan, M., Du, J., and Ward, W.: Detection of migrating diurnal tide in the tropical upper troposphere and lower stratosphere using the Challenging Minisatellite Payload radio occultation data, J. Geophys. Res., 113, D03102, doi:10.1029/2007JD008725, 2008.
Zipser, E. J., Cecil, D. J., Liu, C., Nesbitt, S. W., and Yorty, D. P.: Where are the most intense thunderstorrms on Earth?, B. Am. Meteorol. Soc., 87, 1057–1071, doi:10.1175/BAMS-87-8-1057, 2006.

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13, 1–31, 2013

Impact of land convection

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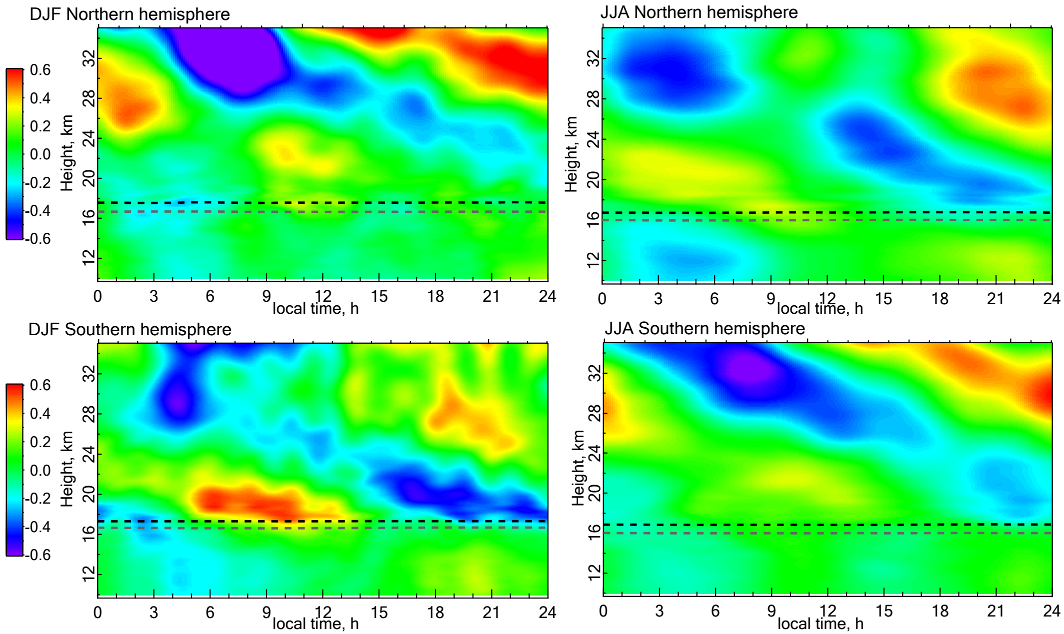

Fig. 1. Zonally-averaged diurnal temperature anomalies in the tropical Northern (top) and Southern (bottom) tropical belts (25° N–0° and 0°–25° S) during DJF (left) and JJA (right). Black and grey dashed lines denote mean CPT and LRT levels, respectively.

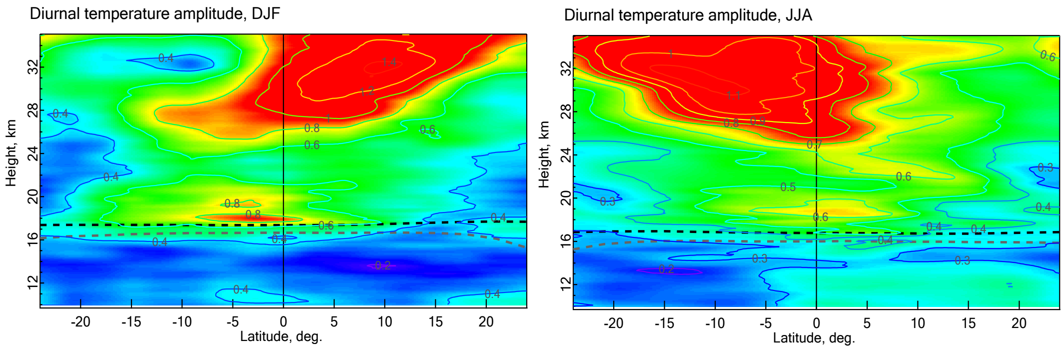

Fig. 2. Zonal-mean amplitude profile of diurnal temperature cycle between 25° N–25° S in DJF (left) and JJA (right).

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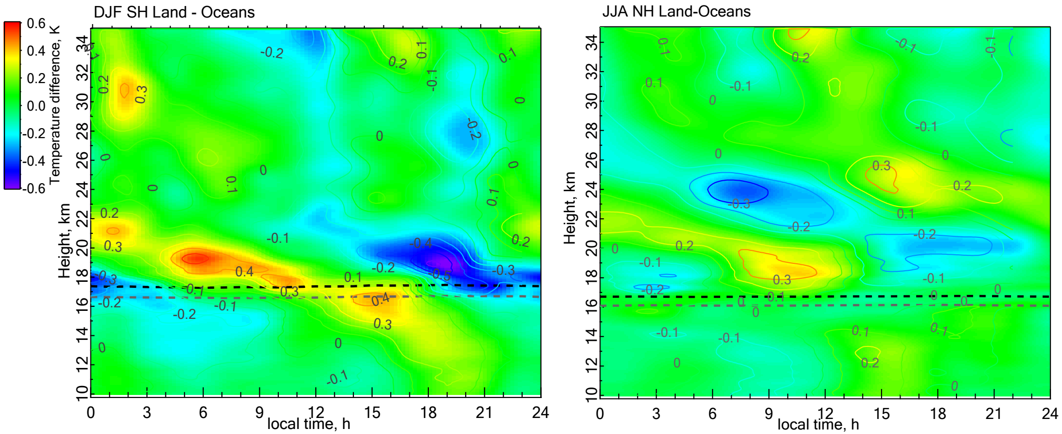

Fig. 3. Difference between vertical profiles of diurnal temperature anomalies above land and oceanic regions in DJF in the SH (left) and in JJA in the NH (right) where land and oceanic domains are defined as areas of highest and lowest TRMM PR OPFs occurrence frequency (Liu and Zipzer, 2005). Black and gray dotted lines denote CPT and LRT levels, respectively above land areas.

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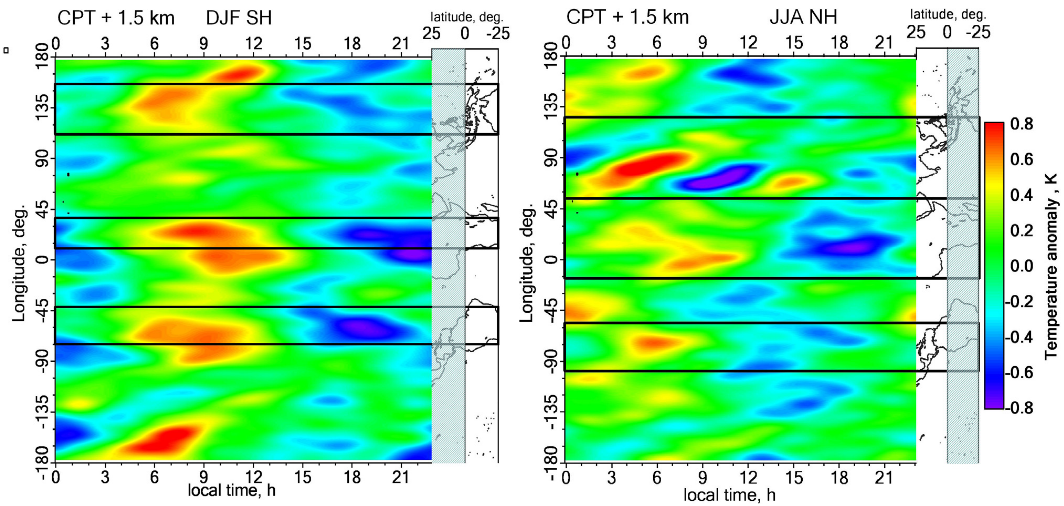

Fig. 4. Geodiurnal variation of diurnal temperature anomaly between 0–25° S and 0–25° N during the local summer in DJF (left) and JJA (right), respectively in a 1.5 km thick layer above the local CPT. Land areas are enclosed in black rectangles, the map on the right giving the details of the geography.

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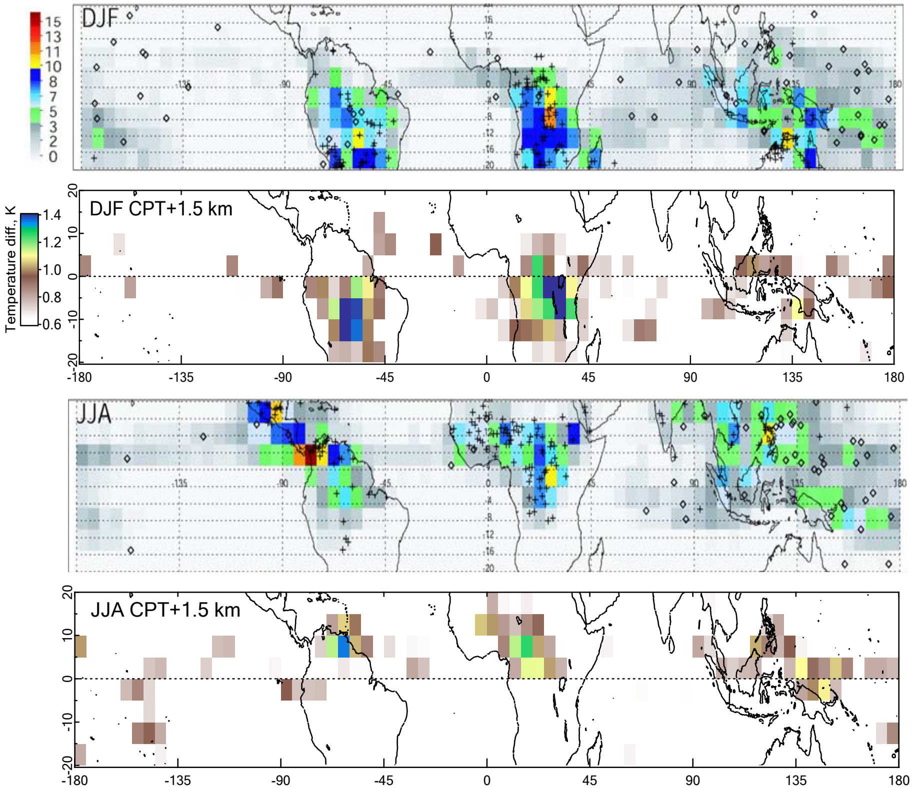

Fig. 5. Comparison between geographical locations of late afternoon cooling above the tropopause (2nd and 3rd panels) and TRMM OPFs density number above 14 km (1st and 3rd panels adopted from LZ05). Top panels: DJF, bottom: JJA. The OPFs frequency of occurrence is in units per thousand in each bin divided by the TRMM 3A25 total pixel number for removing the sampling bias. The color scale of the cooling is the value of the difference between late afternoon (18:00 LT ± 2 h) and mean morning (10:00 LT ± 2 h) temperatures shown in Fig. 3.

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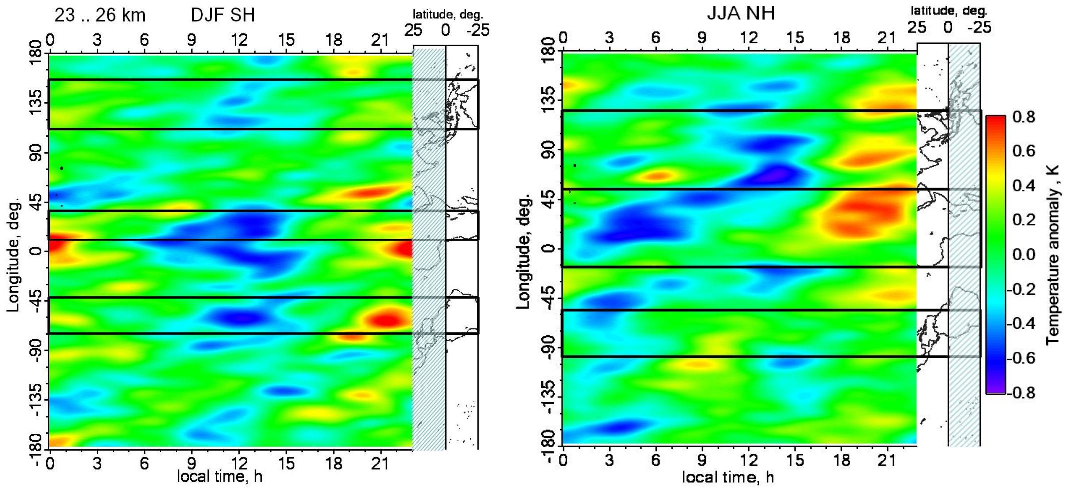

Fig. 6. Geodiurnal variation of diurnal temperature anomaly between 0–25° S and 0–25° N during the local summer in DJF (left) and JJA (right) in a layer between 23 and 25 km altitude. Land areas are enclosed by black rectangles.

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