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Hydration of the lower stratosphere by ice crystal geysers over land convective systems

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Abstract

The possible impact of deep convective overshooting over land has been explored by six simultaneous soundings of water vapour, particles and ozone in the lower stratosphere next to MesoScale Convective Systems (MCSs) during the monsoon season over West Africa in Niamey, Niger in August 2006. The water vapour measurements were carried out using a fast response FLASH-B Lyman-alpha hygrometer. The high vertical resolution observations of this instrument show the presence of enhanced water vapour layers between the tropopause at 370 K and the 450 K level. Most of these moist layers are shown connected with overshooting events occurring upwind as identified from satellite IR images, over which the air mass probed by the sondes passed during the three previous days. In the case of a local overshoot identified by echo top turrets up to 18.5 km by the MIT C-band radar also in Niamey, tight coincidence was found between enhanced water vapour, ice crystal and ozone dip layers indicative of fast uplift of tropospheric air across the tropopause. The water vapour mixing ratio in the enriched layers, up to 8 ppmv higher than that of saturation at the tropopause, and the coincidence with the presence of ice crystals strongly suggest hydration of the lower stratosphere by geyser-like injection of ice particles over overshooting turrets. The pile-like structure of the water vapour seen by the high-resolution hygrometer in contrast to smooth profiles reported by a coarse vertical-resolution satellite observation, suggests that the hydration mechanism described above may be responsible for the known summer seasonal increase of moisture in the lower stratosphere. If this interpretation is correct, hydration by ice geysers across the tropopause may be an important contributor to the stratospheric water vapour budget.

1 Introduction

Water vapour is a key player in stratospheric climate and chemistry. It is the most important greenhouse gas controlling partly the temperature of the stratosphere and upper troposphere (Forster and Shine, 2002), as well as a source of hydroxyl radicals and polar stratospheric clouds involved in ozone depletion (Solomon et al., 1986). Its concentration was reported to increase above Northern America by 5–10% per decade at all levels between 15–26 km over the period 1980–2000 (Oltmans et al., 2000; Rosenlof et al., 2001), whilst a cooling of the lower stratosphere of 0.5 K per decade was reported between 1979–2005 (WMO, 2007), now followed by a levelling off of both parameters since 2001 (Randel et al., 2006). Since troposphere-stratosphere exchange is known to take place primarily in the tropics (Brewer, 1949; Holton et al., 1995), these changes are generally attributed to changes in the vertical transport in the tropical upper troposphere–lower stratosphere (UT–LS). However, the mechanism controlling the amount of water vapour entering the lower stratosphere is still debated. Indeed, whatever is the mechanism of transport invoked – slow radiative ascent, synoptic uplift, or convective overshooting (e.g. Sherwood and Dessler, 2000, and references herein) the amount of water vapour transported is generally assumed to be limited by the minimum temperature of the tropopause resulting in the dehydration of air entering the lower stratosphere (e.g. WMO, 2007). If this mechanism is correct, the observed cooling of the tropopause by 0.5 K per decade since 1979 (Seidel et al., 2001; WMO, 2007), should have resulted in a dryer stratosphere and not the opposite, casting some doubt on the currently accepted process.

Although a number of satellite observations of water vapour in the stratosphere are available, they are of little use in the Tropical Tropopause Layer (TTL) between 14–20 km because of large systematic biases, poor precision, broad vertical resolution and frequent presence of clouds, as shown by Montoux et al. (2007). Most of the information on this layer comes from in situ measurements by sondes, balloons and aircraft, all becoming increasingly available (Kley et al., 1982; Kelly et al., 1993; Jensen et al., 2005; Tuck et al., 2004; Vömel et al., 2002; Weinstock et al., 1995; Richard et al., 2006). However, most of them were carried out over oceanic areas providing little information on land regions known to be preferred location of most intense convective overshooting (Kent et al., 1995, and references therein), and of lightning indicative of stronger updrafts (Vonnegut and Moore, 1958), as confirmed recently by the TRMM (Tropical Rainfall Measuring Mission) radar overshooting precipitation features and the Lightning Imaging Sensor flashes (Liu and Zipser, 2005; Zipser et al., 2006). After the pioneering ER2 total water measurements over Panama in 1980 (Kley et al., 1982) and over Panama and Darwin in 1987 (Kelly et al., 1993), it is only recently that new water vapour profiles over continental convective regions have been made available from the HIBISCUS (Pommereau et al., 2007; Durry et al., 2007) and TROCCINOX (Chaboureau et al., 2007; Corti et al., 2008) balloon and high altitude aircraft campaigns in Brazil and the SCOUT-O3 aircraft deployment in Northern Australia (Vaughan et al., 2008). Remarkably, they are all reporting enhanced moisture layers above the tropopause over land convective systems, never seen in maritime observations where sub-visual cirrus is frequent but limited at and below the tropopause altitude (Kent et al., 1995).

The presence of moist layers in the lower stratosphere over land strongly suggests a hydration mechanism associated with convective overshooting, which could explain also the higher concentration of long-lived species of tropospheric origin such as N_2O and CH_4 in the TTL over land, particularly over Africa, as reported by Ricaud et al. (2007). The explosive convection developing over land in the afternoon could lead to strong overshooting of adiabatically cooled air above the cold point tropopause (CPT) in the stratosphere as shown by Danielsen (1993), resulting in a systematic cooling of the lower stratosphere in the afternoon as shown by Pommereau and Held (2007). It could also explain the presence of ice particles as suggested by Liu and Zipser (2005) from the relationship between precipitating ice mass and the mass of small ice particles seen by the TRMM radar, eventually evaporating, thus hydrating and not dehydrating the lower stratosphere as proposed by Danielsen (1982). The injection of ice particles above CPT up to 18–19 km over intense land convective systems was confirmed recently by the balloon and aircraft campaigns over Brazil and Northern Australia (Nielsen et al., 2007; Chaboureau et al., 2007; Corti et al., 2008). In the case of Brazil these events were fully reproduced by Cloud Resolving Models (Chaboureau et al., 2007;

Grovesnor et al., 2007) suggesting a local injection of a few tons of water per second in the lower stratosphere. As reported by in-situ aircraft measurements of H_2O and HDO (Hanisco et al., 2007), similar events may also occur in the extratropical stratosphere displaying isotopic water signatures that are characteristic of sublimated ice lofted from the troposphere during convective storms.

Aiming at better exploration of this possible mechanism of hydration of the lower stratosphere by convective overshoots, a series of simultaneous water vapour, particle and ozone measurements has been carried out within a new SCOUT-AMMA campaign in August 2006 from Niamey, Niger (13.6° N) in West Africa, a location of frequent overshooting during the monsoon season according to Liu and Zipser (2005). Indeed, the summer season there is characterized by powerful convective systems, commonly known as MCS (Mesoscale Convective Systems), developing due to the confrontation of moist air from the Gulf of Guinea carried by the monsoon and dry air from the Saharan heat low carried by the Harmattan wind. Those MCSs developing initially over East and Central Africa are travelling westwards and their convective activity is strongly modulated by the diurnal cycle with a maximum development around 16 local time (e.g. Liu and Zipser, 2005).

Within the SCOUT-AMMA campaign named after the support of two projects of the European projects SCOUT-O3 (http://www.ozone-sec.ch.cam.ac.uk/scout_o3/) and AMMA (http://www.amma-international.org) a number of soundings (29) of various types and larger balloon flights (7) carrying a variety of instruments were performed, moreover in coordination during the first half of the month with high altitude M-55 Geophysica aircraft flights from Ouagadougou, Burkina-Faso, 400 km West of Niamey. Among the sondes, six of them were carrying integrated packages of water vapour, ozone and particle measuring instruments. These sondes were flown as close as possible to MCSs identified on Meteosat Second Generation (MSG) and MIT C-band radar images (Williams et al., 2008). The instruments flown on each were a FLASH-B Lyman-alpha hygrometer (Yushkov et al., 1998), a backscatter sonde (Rosen and Kjome, 1991), an Electro-Chemical Cell (ECC) ozonesonde and a Vaisala RS-92 meteorological package (pressure, temperature, GPS location and altitude). The sondes were reaching 30 km altitude, after which the measurements were continued during the fast descent below a parachute. The connection with possible overshoots upwind was analyzed by examination of MSG images available every 15 min along backward trajectories using the method of brightness temperature difference between 6.2 and 10.8 μm for detecting the overshoot signatures (Schmetz et al., 1997) and cloud top information provided by the C-band radar operated with continuous 10-min volume scans.

The organisation of the paper is the following. Section 2 provides a description of the instruments and details of the experiment, whose results are presented in Sect. 3. The impact of overshooting systems on the lower stratospheric water content and the relation between ice particles and water vapour over a local convective system are discussed in Sect. 4, followed by concluding remarks in Sect. 5.

2 Experimental setup

The payload for the sondes lifted by an Aerostar 1500 m^3 plastic balloon, included a backscatter (BKS) sonde, an EnSci Electro Chemical Cell (ECC) ozone sonde, a Vaisala RS-92-SGP PTU radio sonde (H-Humicap), and a FLASH-B hygrometer coupled with a Vaisala RS-80 radiosonde at the bottom of the flight train 55 m below the balloon, and a parachute, for a total weight of 10 kg, allowing ascent to 30 km. The BKS sonde (see Rosen and Kjome, 1991, for technical details) is measuring the backscatter coefficients at two wavelengths (940 nm and 480 nm) by sending out a light flash approximately every 7 s, and counting the backscattered photons with two photodiodes.

The FLASH-B (FLuorescence Advanced Stratospheric Hygrometer for Balloon) instrument is the Lyman-alpha hygrometer developed at the Central Aerological Observatory for balloon-borne water vapour measurements in the upper troposphere and stratosphere (Yushkov et al., 1998, 2001). The instrument is based on the fluorescent method (Kley and Stone, 1978; Bertaux and Delannoy, 1978), which uses the photodissociation of H_2O molecules at a wavelength <137 nm followed by the measure-

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ment of the fluorescence of excited OH radicals. The source of Lyman-alpha radiation (λ=121.6 nm) is a hydrogen discharge lamp, while the detector of OH fluorescence at 308–316 nm is a Hamamatsu R647-P photomultiplier run in photon-counting mode with a narrow band interference filter selecting the fluorescence spectral region. The intensity of the fluorescent light sensed by the photomultiplier is directly proportional to the water vapour mixing ratio under stratospheric conditions (10–150 hPa) with small oxygen absorption (3% at 50 hPa). The H_2O measurement range is limited to pressures lower than 300–400 hPa due to strong Lyman-alpha absorption in the troposphere.

The instrument uses an open optical layout design (Khaplanov et al., 1992), where the optics is looking directly outside. Such arrangement is suitable for nighttime measurements only at solar zenith angles larger than 98°. The co-axial optical layout allows a reduction in the size of the instrument to 106×156×242 mm for a total weight of about 1 kg including batteries. The hygrometer is coupled with a Vaisala RS-80 radiosonde, providing telemetry as well as pressure and temperature measurements.

Each FLASH-B is calibrated in the laboratory before flight. A description of the procedure can be found in Vömel et al. (2007). The precision of the measurements is 5.5% for a 4-s integration time at stratospheric conditions, while the accuracy is limited by the calibration error estimated at 4% in the 3–200 ppmv range. The total uncertainty is less than 10% at the stratospheric mixing ratios greater than 3 ppmv, increasing to about 20% at mixing ratios less than 3 ppmv. Accuracy and good performance of the FLASH-B instrument have been confirmed through point-by-point comparisons with the NOAA-CMDL frost point hygrometer during the LAUTLOS-WAVVAP intercomparison campaign (Vömel et al., 2007) showing excellent agreement between both instruments with a mean deviation of –2.4±3.1% (1 standard deviation) for data between 15 and 25 km altitude.

While the minimum response time of FLASH-B instrument is 0.2 s, here we will use the data averaged over 4 s, resulting in a vertical resolution of 20 m during ascent and 100 m during the fast descent in the stratosphere. The altitude used in this study is that of the GPS, and the potential temperature is that calculated from pressure and temperature provided by the Vaisala RS-92 radiosonde.

The flight configuration of FLASH-B, in which the analyzed volume is located beneath the downward looking optics about 24 mm away from the lens, caused noticeable self-contamination due to water outgassing from the instrument surfaces during the ascent in the stratosphere above 90 hPa. In this arrangement, the measurements during the fast descent in undisturbed air can be considered as contamination-free as shown by the drop of water vapour immediately after the balloon burst.

3 Flight results

The data available are those of the six soundings, all performed in the evening after sunset and downwind or next to MCSs. The details of the six flights are shown in Table 1. The water vapour, backscatter and temperature profiles of each are displayed in Fig. 1 and the average water vapour profile along with the standard deviation are shown in Fig. 2.

The cold point tropopause (CPT), the minimum temperature, is found at 16.5 km on average with a mean temperature of –79.5°C, varying between 16.3 km and 16.8 km and with temperatures ranging from –77.5°C to –81.5°C. This is significantly lower and warmer than the 17–17.5 km, –85/–90°C CPT over Darwin. The BKS sonde indicates the frequent presence of cloud anvils and cirrus in the upper troposphere sometimes up to the CPT, a permanent layer of aerosols between 19–21 km, and sporadic thin layers of particles in the lower stratosphere, particularly enhanced on 23 August. The aerosol layer at 19–21 km seen also by all other balloon and aircraft particle measuring instruments during the campaign (optical particle counter, UV-Vis solar occultation spectrometer, backscatter diode laser) was made of 0.1 μm radius, undepolarizing and thus spherical liquid particles, the characteristics of volcanic aerosols. Since they could be observed also by the CALIPSO lidar from the beginning of the satellite operation in June 2006, they are attributed to the eruption of Soufriere Hills on Montserrat Island in the Caribbean on 20 May 2006 (Vernier et al., 2007). The thin particle layers at 17.5–

18 km above the tropopause are very similar to those reported in Brazil by the same backscatter sonde up to 20.8 km (Nielsen et al., 2007), and by the Forward Scattering Spectrometer Probe (FSSP)-100 on board the M-55 Geophysica aircraft at around 18 km (Chaboureau et al., 2007). Since they could be seen during descent below a parachute they cannot be attributed to particles falling from the balloon. The colour index of the BKS sonde, i.e. the ratio between the aerosol backscatter ratios at 940 nm and 480 nm, and the depolarized signal of the backscatter diode laser flown on another balloon indicate that, in contrast to the volcanic aerosols at 20 km, these particles are non spherical and of 0.5–10 μm radius, the characteristics of small ice crystals.

The FLASH-B relative humidity (RH) in the troposphere in the 350 hPa–150 hPa range show very good agreement with the Vaisala RS-92 H-Humicap RH (mean difference of -0.27 ± 6.15% RH). At altitude below 350 hPa FLASH-B is incapable of measuring humidity due to high absorption of Lyman-alpha emission by water vapour. In the stratosphere FLASH-B data, showing good repeatability, are also consistent with the measurements by the fast in situ Lyman-alpha hygrometer (FISH) (Zoger et al., 1999) flown on the high-altitude M55 Geophysica aircraft during the same period from Ouagadougou, displaying a minimum mixing ratio of 4.2 ppmv at 19 km (450 K) (C. Schiller, personal communication, 2007), compared to the 4.4 ppmv value seen by FLASH-B at the same level. The profiles of the micro-SDLA tunable diode laser hygrometer (Durry et al., 1999) flown on a larger balloon in parallel with the sondes on 5 and 23 August showed similar profiles in the TTL, but for an unknown reason biased systematically low by a factor 1.8 and 2, resulting in doubtful values of 2.6 ppmv and 2.1 ppmv at 18.3 and 20 km, respectively, in the two flights.

Overall, the FLASH-B water vapour profiles display an almost invariably saturated or supersaturated upper troposphere and an average mixing ratio (MR) of 6.5 ppmv at the tropopause. A broad MR minimum of 4.1–4.4 ppmv, the hygropause, is centred at around 20 km, that is about 3.5 km above the cold point tropopause, surmounted by an almost constant amount of 5.7 ppmv in the stratosphere above 24 km varying very little (0.2 ppmv standard deviation) from one flight to another. In contrast to these levels,

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the water vapour amount is highly variable between the CPT (near 16.5 km, 365 K) and 19 km (450 K), a region displaying frequently enhanced water vapour layers.

In the upper troposphere the water vapour profiles show saturation or supersaturation with respect to ice, except on 5 August when the CPT temperature was as high as −77.5°C and no clouds were observed by the BKS sonde. The air inside thick cirrus clouds or anvils is always saturated or supersaturated. Supersaturation in cloud-free conditions is frequently observed too. On 14, 21 and 23 August cirrus anvil cloud tops reach the CPT level according to the BKS sonde measurements. A maximum saturation ratio as high as 177% RHi is observed at 15.9 km a little below the CPT in a cumulonimbus anvil on 14 August. No saturation was observed higher than 600 m above the CPT.

A better insight into the vertical structure of water vapour in the stratosphere is provided by Fig. 2 showing the mean water vapour profile calculated from the six soundings along with the standard deviation. Above the minimum at 475 K (20 km), the mean mixing ratio is gradually increasing to a maximum of 5.8 ppmv around 620 K (24.7 km), above which the mixing ratio is almost constant or slightly decreasing up to 30 km. The observed structure is very consistent with the water vapour tape recorder signal derived from HALOE by Mote et al. (1996). It corresponds to an uplift of the minimum mixing ratio from about 17–17.5 km during the season with the coldest tropopause temperature in the Northern Hemisphere winter in January–February at a vertical velocity of 0.5 km/month (0.2 mm/s) consistent with the Brewer-Dobson circulation. However, between the CPT and the hygropause, the picture provided by the high-resolution measurements differs significantly from the satellite profiles with low vertical resolution. The water vapour increase in this layer in the Northern Hemisphere summer season does not show a smooth profile from 6.5 ppmv at the CPT to 4.2 ppmv at the hygropause, which would be the case if it was due to a slow uplift of well mixed moist air from the warmer tropopause. Instead, it appears as an accumulation of thin layers, a few hundred meters thick, suggesting relatively fresh successive injections of water at variable altitudes up to 19 km (450 K). Moreover, the water vapour enhancements with respect to the mean profile in those layers are highly variable, ranging from 0.2 ppmv (3 August) to 7 ppmv (23 August). Indeed, the standard deviation (Fig. 2), accounting for the variability between the profiles, decreases to a quasi-constant value of 0.2 ppmv only by 450 K (19 km), implicitly suggesting that the lower stratosphere is influenced by a succession of convective overshootings, carrying moisture across the tropopause. However, the amplitude of hydration remains far from saturation except when immediately above the CPT. Among all profiles that of 3 August, closer to the beginning of the convective season is the driest with a minor enhancement at 17.2 km, whilst most of the others are displaying multiple layers between the CPT and 20.5 km (5 August), the largest of 13 ppmv being observed 0.8 km above the CPT on the last flight on 23 August.

4 Discussion

4.1 Impact of convective overshoots

A key question is how fresh each of these water vapour layers seen in the lower stratosphere are and whether they relate to overshoots upwind. This question was addressed by looking at overshoots, which might have occurred upwind along backward trajectories ending at the sounding location.

The trajectories are three-day backward 3-dimensional trajectories calculated every 5 K in the 350–450 K range using the TRACAO trajectory model (Lukyanov et al., 2003), based on ECMWF T511 operational analyses (21 pressure levels). The calculations are performed by time steps of 15 min corresponding to the time resolution of MSG satellite images, a product of the SEVIRI (Spinning Enhanced Visible and Infrared Imager) instrument operating onboard the Meteosat-08 (MSG-01) satellite. The presence of convective overshoots along the trajectories is identified from MSG images by the brightness temperature difference (BTD) technique (Schmetz et al., 1997). The method is based on the observation of the difference between cloud top emissions at two wavelengths, 6.2 μm and 10.8 μm, the first being sensitive to the water vapour emission at higher temperature in the lower stratosphere above the cloud in contrast to the adiabatically cooled turret. Following Chaboureau et al. (2007), overshoots are declared when the BTD exceeds 3 K within a MSG pixel of 3 km×3 km, although individual turrets may be substantially smaller than this. Overshoots are detected along each backward trajectory together with an estimate of their size given by the number of MSG pixels with BTD larger than 3 K.

Figure 3 shows an example of backward trajectories at different potential temperature levels starting 30 h prior to the sounding at 20:00 UTC on 5 August and superimposed on an MSG BT image. The two trajectories ending at 395 K and 400 K passed over an overshooting MCS at 14:00 UT on 4 August, i.e. 30 h prior to the sounding. The time when the air mass was passing over an overshoot area and the size of the latter were explored for each trajectory. An example of the results of such analysis for 5 August is shown in Fig. 4. The time at which overshoots were encountered are represented by markers at the altitude of the water vapour-enhanced layers of size proportional to the number of MSG pixels of BTD>3 K. In this example, three overshoots were encountered, a large one of 23 pixels and a smaller one of 12 pixels around 390 K 30 h before the sounding, and a small one of 6 pixels at 410 K 48 h before. The three are corresponding to water vapour layers centred at 17.1 km (391 K) and 17.9 km (411 K). However, no overshoot could be identified corresponding to the layer at 20.3 km (492 K).

The information relevant to all soundings is summarized in Table 2, including an estimate of the vertical velocity of the overshoot above the equilibrium level assumed at 14 km, using the rule of thumb (Vonnegut and Moore, 1958). A connection with an overshoot upwind during the three previous days could be identified for 70% of the layers, occurring between 16:00 h local in the afternoon and 04:00 h during the night but with no marked preferred time. Ignoring the specific case of the large 7 ppmv layer on 23 August, which will be treated in more detail later, the amplitude of water vapour enhancements varying between 0.2 to 1.8 ppmv correlates roughly linearly with the size of the overshoots. With the exception of the layer at 20.3 km on 5 August for which no overshoot could be found, the required velocity of the updraft above the equilibrium level is between 50–80 m/s according to the rule of thumb proposed by Vonnegut and Moore (1958). An order of magnitude of such vertical velocity as well as that of the water vapour enhancements are consistent with those suggested by cloud resolving models (Chaboureau et al., 2007; Grovesnor et al., 2007).

However, no direct connection with an overshoot could be identified for about 30% of the layers. The possible reason for that could be that the overshoot occurred more than 72 h before the sounding over Ethiopia or possibly also over India during the monsoon season, East of 40° E , the limit of the MSG imagery used in this study.

In summary, the unambiguous connection between enhanced water vapour layers reported by the sondes over Niamey and earlier upwind overshoots incorporates the idea of a hydration process of the lower stratosphere by fast uplift of water condensates across the tropopause over continental areas. Although such events have been already observed in the southern tropics during the local summer (Nielsen et al., 2007; Chaboureau et al., 2007), they seem to be particularly abundant in the northern tropics during the monsoon season in Africa and Asia. The larger geographical extension of continental convective regions in the Northern Hemisphere (Liu and Zipser, 2005) may explain the higher concentration of water vapour in the lower stratosphere during the summer compared to the same season in the Southern Hemisphere.

4.2 Ice particles and water vapour over a local convective system on 23 August 2006

The next step is to understand how water amounts larger than those permitted by condensation at the temperature of the tropopause could penetrate the stratosphere. Three out of six profiles (14 August, 21 August and 23 August) show the presence of thin layers of particles above the CPT similar to that reported up to 20.8 km over Brazil by Nielsen et al. (2007). Among these, the largest are observed on 23 August. An enlargement of the latter is shown in Fig. 5. The plot also includes the ozone profile measured during ascent and advanced by 50 s to correct for the ECC sensor time constant, which corresponds to a vertical displacement of about 250 m.

In addition to the existence of upwind overshoots already identified, the specificity of this sounding in contrast to all others is that it was launched within a broad convective area and particularly two hours after the pass of a small-size convective system over the launching station when the co-located C-band radar was showing turrets developing up to 18.5 km (Fig. 6). The most remarkable features in the plot in Fig. 5 are: a) ice saturation in a cloud anvil expanding from 13–16 km and saturating the BKS signal; b) a relative minimum of water vapour MR at the cold point tropopause of −78°C at 15.8 km corresponding to saturation; c) the presence of a strongly supersaturated (138% RHi) sharp water vapour peak immediately above at 16.2 km, also present in the ascent profile (not shown) and thus not an instrumental artefact, coincident with an ice crystals layer composed of 0.5–10 μm size particles, according to the BKS colour index; and d) the presence of small water vapour-enhanced layers up to 18.5 km in a subsaturated region also coincident with layers of ice crystals, but on the descent profile only and not during the ascent an hour earlier, 50 km East.

Remarkable also is the presence of dips in the ozone profile coincident with the water vapour and ice crystals layers, indicative of an injection of relatively ozone-poor tropospheric air into the lower stratosphere. Although low-biased by a factor 2 compared to that of FLASH-B, the water vapour profile reported by the micro-SDLA tunable diode laser instrument flown on a larger balloon at 18:15 UT, an hour before the sondes, shows a similar 1 km broad peak of 6 ppmv at 16 km above a minimum of 3.8 ppmv at the tropopause, at 15.2 km only. Note that its lower altitude compared to all other flights might be coming from the injection of adiabatically cooled air over the local convective system which, as shown by Pommereau and Held (2007), is resulting in the lowering of the CPT. Also flown on the same balloon, the LABS (Laser Backscatter Sonde) instrument shows several layers of depolarizing and thus solid particles at 19.2 km, but in that case during the ascent of the balloon and not during its slow descent 2 h later.

Overall, all balloon measurements confirm the sporadic presence of short-lived layers of ice crystals above the tropopause over a convective area, coincident with observations of radar echo tops displaying turrets penetrating deeply into the stratosphere, as well as dips in ozone profiles at the same levels indicative of fast uplift of tropospheric air. Those observations provide experimental evidence of hydration of the stratosphere after sublimation of ice crystals freshly injected by overshooting turrets, a geyser-like feature, as shown by pictures of severe storms taken from a Learjet (Fujita, 1992) and suggested by cloud resolving model simulations (Chaboureau et al., 2007; Grovesnor et al., 2007).

5 Concluding remarks

A series of six balloon flights combining a fast response FLASH-B Lyman-alpha hygrometer and particles and ozone measurements supported by C-band ground-based radar observations has been carried out in West Africa during the monsoon season. The high vertical resolution observations show evidence of the presence of local accumulations of water vapour-enhanced layers between the tropopause at 370 K (16.0 km) and the 450 K level (19 km). Most of them are shown connected with overshooting events upwind identified from satellite IR images, flown over by the air mass probed by the sondes along three-day backward trajectories. In the case of a local overshoot identified by echo tops turrets up to 18.5 km in the C-band radar, tight coincidence was found between enhanced water vapour, ice crystal and ozone dip layers indicative of fast uplift of tropospheric air across the tropopause. The water vapour mixing ratio in the enriched layers, higher than that of condensation at the tropopause, and the coincidence with the presence of ice crystals strongly suggest hydration of the lower stratosphere by geyser-like injection of ice particles over overshooting turrets.

The local accumulation of water vapour seen by the high-resolution hygrometer in contrast to smooth profiles reported by broad resolution satellite observations, suggests that the above hydration mechanism may be responsible for the known summer seasonal increase of moisture in the lower stratosphere. The mechanism offers also an explanation for the contradiction between the increase of moisture in the stratosphere between 1980–2000 and the concomitant cooling of the tropopause, which would result in the opposite behavior if the dehydration process proposed by Danielsen (1982) was correct. Ice geyser hydration across the tropopause may be a significant factor controlling water vapour in the stratosphere on a global scale.

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Table 1. List of soundings (Local Time=UTC+1).

Date	Launch time, UTC	Time interval of descent UTC	Float altitude (m)
3 Aug	18:40	20:23–20:46	29 970
5 Aug	18:58	20:40–21:13	30 146
7 Aug	18:40	20:38–21:16	30 978
14 Aug	19:47	21:25–21:55	29 006
21 Aug	21:53	23:44–00:08	30 710
23 Aug	19:07	20:39–21:13	25 987

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Table 2. Summary of observed water vapour-enhanced layers and correlated convective overshoots upwind. From left to right: date of sounding, altitude and potential temperature of the layer, amplitude of enhancement compared to background, thickness of the layer in altitude and potential temperature units, delay between the overshoot and the sounding, local time of overshoot, size in MSG pixels and vertical velocity at the equilibrium level at 14 km estimated as in Vonnegut and Moore (1958).

Date	H_2O layer altitude (km) pot. temp. (K)	H_2O enhancement (ppmv)	Layer thickness (km/K)	Time since overshoot (h)	Local time of overshoot, (h)	Size of overshoot (pixels)	Vertical velocity (m/s)
3 Aug	17.2/395	0.2	0.6/10	22	23	5	64
5 Aug	17.1/391	1.5	1.2/30	30	15:15	23+12	62
5 Aug	17.9/411	0.6	0.6/9	47	23:30	6	78
5 Aug	20.3/492	0.6	1/16	n/a	n/a	n/a	126
7 Aug	17.4/390	0.3	0.5/17	n/a	n/a	n/a	68
14 Aug	16.9/377	1.8	0.4/14	21, 19	01, 03	75+65	58
14 Aug	17.0/384	1.0	0.7/19	44	03	33	60
14 Aug	17.3/394	1.3	0.7/19	n/a	n/a	n/a	66
14 Aug	18.2/416	0.7	0.7/19	n/a	n/a	n/a	84
21 Aug	16.6/378	0.8	0.3/11	5	19	11+12	52
21 Aug	16.9/390	0.2	>0.2/>0.7	6	18	6	58
21 Aug	17.2/408	1.0	0.7/24	6	18	7+3	64
23 Aug	16.1/368	7.0	1.2/21	19, 23	03, 23	7+7+6	62
23 Aug	17.6/392	1.1	0.3/5	22	00	5	72
23 Aug	18.3/420	0.5	0.3/9	19	03	80	8.6

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Hydration of the lower stratosphere by ice crystal geysers

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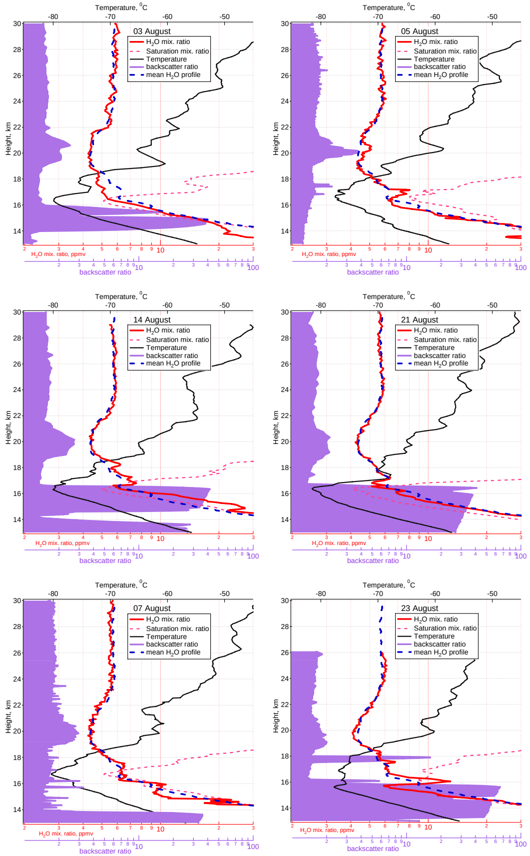

Fig. 1. Results of the six soundings carried out from Niamey in August 2006. Water vapour mixing ratio during descent (red solid), backscatter ratio at 940 nm during descent, except ascent on 3 and 5 August (violet filled to zero), temperature (RS80 descent, solid black), mean water vapour of the six soundings, (dashed blue) and saturation mixing ratio (dashed pink). The dates are shown on each plot.

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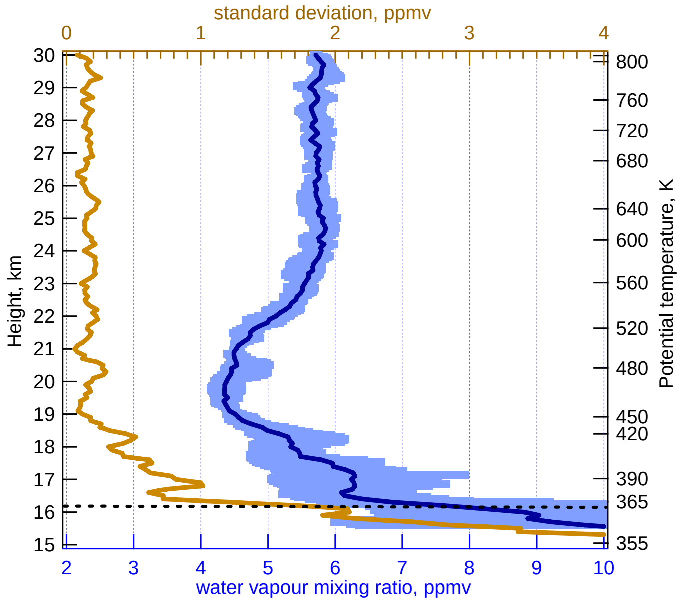

Fig. 2. Mean water vapour profile calculated from the six soundings (blue, bottom axis) and standard deviation (brown, top axis). The blue shaded area embraces the range of measured water vapour profiles.

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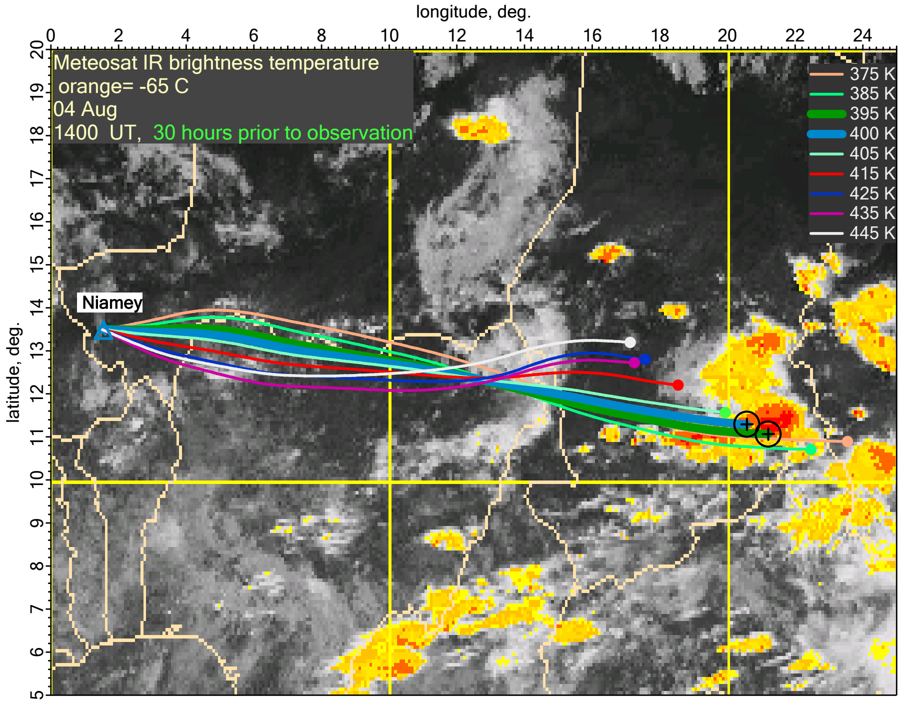

Fig. 3. 30-h backward trajectories ending between 375 and 445 K at the sounding location at 20:00 UT on 5 August superimposed on the MSG 10.8 μm brightness temperature image at 14:00 UT on 4 August. The location of air parcels at each level at the time of the MSG image is indicated by thick dots. The trajectories at 395 K and 400 K passing by that time over an MCS are represented by thicker lines and black-circled crosses.

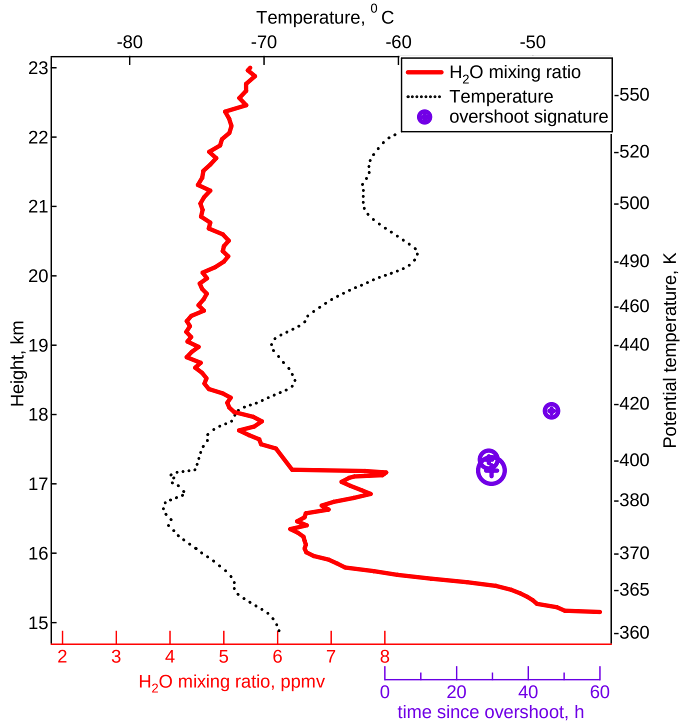

Fig. 4. Water vapour (solid red, bottom axis) and temperature (dotted black, top axis) profiles above Niamey on 5 August 2006. The violet markers plotted against a secondary bottom axis indicate backward trajectories, which have passed over a convective overshoot identified in MSG satellite images by the BTD technique (see text). The secondary bottom axis shows the time delay between the overshoot and the sounding. The markers are size-coded with respect to the number of MSG overshooting pixels.

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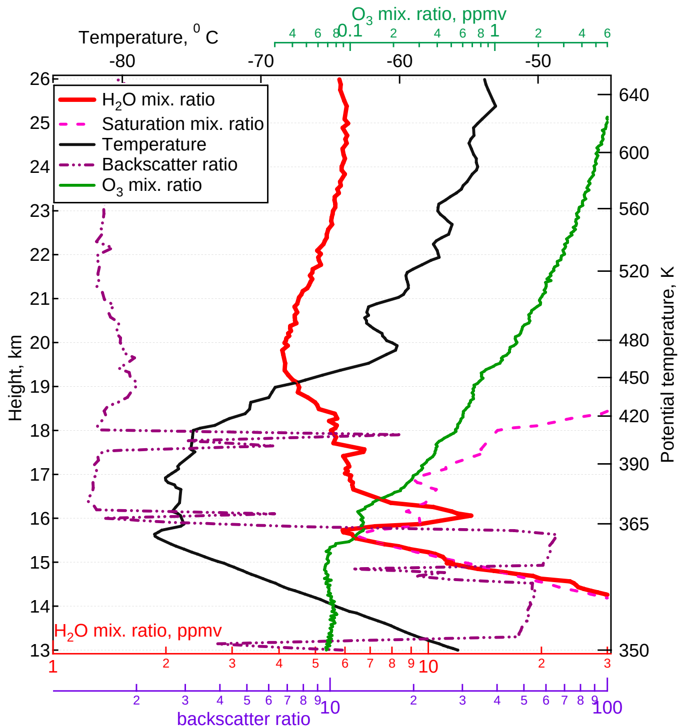

Fig. 5. Vertical profiles of water vapour mixing ratio during descent (solid red), RS-92 descent temperature (solid black), saturation mixing ratio (dotted magenta), backscatter ratio at 940 nm during descent (dash-dotted violet) and ozone mixing ratio during ascent (solid green) on 23 August 2006.

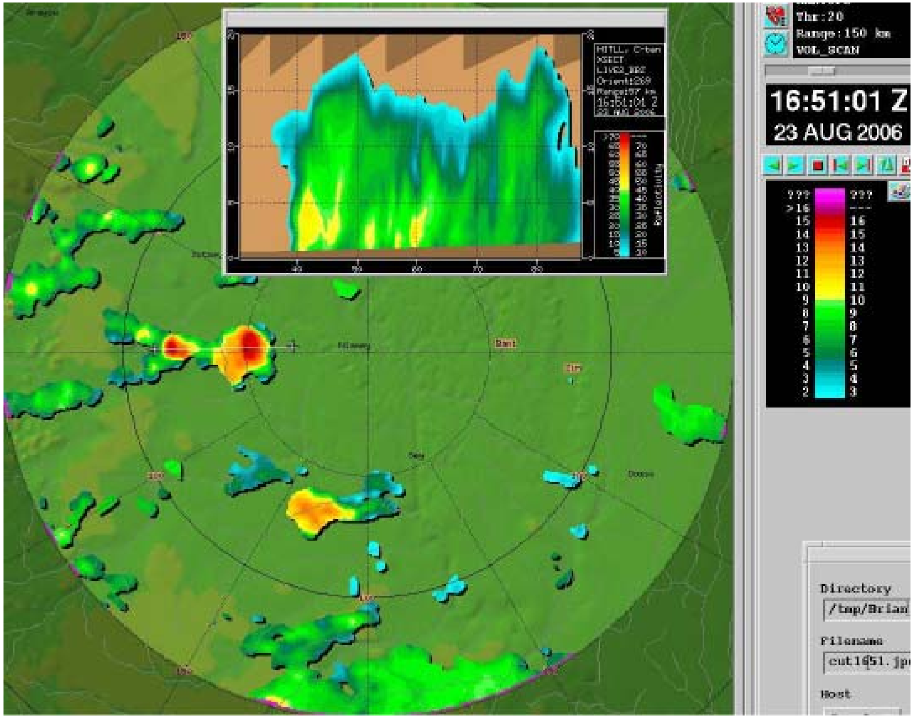

Fig. 6. MIT C-band radar cloud top heights at 16:51 UT on 23 August, two hours before the sounding from Niamey: **(a)** (large plot) geographical distribution of cloud tops within 150 km range colour coded by the cloud top altitude (colour code on top right scaled from 2 km in light blue to >16 km in magenta with 1 km step); **(b)** (small plot) altitude-distance cross-section of the thunderstorm between 30–90 km west of Niamey with altitude markers at 0, 5, 10, 15 and 20 km. The sonde launched at 19:07 was heading true West and therefore probed the lower stratosphere downwind of the thunderstorm.