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Abstract

Water vapor sonde observations were conducted at Bandung, Indonesia (6.90° S, 107.60° E) and Tarawa, Kiribati (1.35° N, 172.91° E) in December 2003 to examine the efficiency of the "cold trap" dehydration in the tropical tropopause layer (TTL). Trajectory analysis based on bundles of trajectories suggest that the modification of air parcels' identity due to irreversible mixing by the branching-out and merging-in of nearby trajectories is found to be an important factor, in addition to the routes air parcels are supposed to follow, for interpreting the water vapor concentrations observed by radiosondes in the TTL. Clear correspondence between the observed water vapor concentration and the estimated temperature history of air parcels is found showing that dry air parcels are exposed to low temperatures while humid air parcels do not experience cold conditions during advection, in support of the "cold trap" hypothesis. It is suggested that the observed air parcel retained the water vapor by roughly twice as much as the minimum saturation mixing ratio after its passage through the "cold trap," although appreciable uncertainties remain.

1 Introduction

Water plays a crucial role on the earth's radiation budget through cloud formation and greenhouse effect. Although its direct anthropogenic production is negligible, globally averaged evaporation from the surface are projected to increase in response to global warming (Houghton et al., 2001), leading to an increase in the tropospheric water vapor concentration. In the lower stratosphere, there is a cooling trend (Randel and Cobb, 1994; Ramaswamy et al., 2001) as a consequence of the decrease in the ozone concentrations (Ramaswamy et al., 1996; Hare et al., 2004) and possibly the increase in stratospheric water vapor (Forster and Shine, 1999). The water vapor increase from the early 1980's to around 2000 amounts to about 1% per year (Oltmans and Hofmann, 1995; Evans et al., 1998), which is much greater than that estimated from the

observed rise of methane concentration through its oxidation (Kley et al., 2000). The stratospheric temperature decrease could cause a delay in the expected ozone recovery due to the ban of chlorofluorocarbons (Shindell et al., 1998). Since water takes part in the ozone chemistry through the formation of hydroxyl radical and polar stratospheric clouds (PSCs), the anthropogenic ozone loss will be enhanced if stratospheric concentration increases. Modification of stratospheric temperatures could lead to a modulation of transport properties which could further influence the stratospheric ozone and water vapor distributions on a global scale. Little is known about the complex feedback processes in the radiation-dynamics-chemistry of the troposphere-stratosphere system.

A basic understanding of water vapor in the stratosphere starts with the idea that it must reflect the temperature history the air experienced before entering the stratosphere. This, together with the global distribution of atmospheric ozone, lead to the description of the general circulation in the stratosphere in which air must enter the stratosphere through the cold tropical tropopause and descend in high latitudes of both hemispheres (Brewer, 1949). Newell and Gould-Stewart (1981) proposed the "stratospheric fountain" hypothesis in which the entry of tropospheric air must be restricted to the tropical western Pacific during boreal winter and the Bay of Bengal in summer. This hypothesis was generally accepted until Sherwood (2000) indicated that the vertical motion near the tropopause over the western Pacific is downward. This finding was supported in numerical simulations (Gettelman et al., 2000; Hatsushika and Yamazaki, 2001).

Big conceptual changes took place in the 1990's. The upward motion in the tropical lower stratosphere was found to be driven by extratropical wave drag (Haynes et al., 1991; Holton et al., 1995; Plumb and Eluszkiewicz, 1999). Evidence of dehydration associated with passage through the tropical tropopause has been shown to be imprinted in the vertical profile of water vapor mixing ratio (Mote et al., 1996). Highwood and Hoskins (1998) shed light on the idea of the tropical transition layer (or tropical tropopause layer; TTL) introduced by Atticks and Robinson (1983). The tropical tropopause is now no longer a clearly defined boundary between the troposphere

and the stratosphere but instead should be treated as a transition layer extending from around 200 hPa to 80 hPa levels. This layer is located above the reach of tropospheric deep convection but below the tight control of the extratropical suction pump, thus dividing the two major dynamical processes that could affect the tropical tropopause.

The introduction of the TTL prompted a novel idea on the dehydration processes in the tropical tropopause region. Holton and Gettelman (2001) proposed a new mechanism of dehydration by using a simple mechanistic model, in which the air parcels are dehydrated during the horizontal advection through the "cold trap" region. This hypothesis exhibits a strong contrast to earlier views in which the cooling due to vertical motion plays an essential role in dehydration. This hypothesis was further examined by using a trajectory model (Gettelman et al., 2002), GCM simulations (Hatsushika and Yamazaki, 2003), and satellite water vapor data (Randel et al., 2001). It is now becoming widely accepted as a key dehydration process for tropospheric air entering the stratosphere.

It should be worth mentioning, however, that the TTL is not a calm region but it is full of disturbances generated by atmospheric waves (Tsuda et al., 1994; Fujiwara et al., 1998). These waves are shown to work as a "dehydration pump" that introduces dry stratospheric air into the uppermost troposphere on the one hand and blocks the entry of humid tropospheric air into the stratosphere on the other (Hasebe et al., 2000; Fujiwara et al., 2001). The variabilities with the intraseasonal oscillation (ISO) could also affect the dehydration efficiency of the TTL. Eguchi and Shiotani (2004) have shown that the "cold trap" dehydration does not occur continuously but rather takes place intermittently in an organized system that takes the form of the combined Kelvin- and Rossby-wave response to the thermal forcing. It is thus strongly controlled by the life cycle and the passage of the ISO. TTL is also subject to hydration due to the northern summer subtropical monsoon. There is evidence that the transport across the subtropical tropopause is strongly influenced by the upper-level anticyclones associated with the monsoons (Chen, 1995; Postel and Hitchman, 1999; Dethof et al., 1999). As the monsoon anticyclones extend up into the stratosphere, they could also contribute

to tropical-extratropical exchange in the lower stratosphere (Horinouchi et al., 2000; Gettelman et al., 2004b).

We are now aware that a large variety of processes that are mutually dependent upon each other through complicated feedbacks are responsible for determining the climatology and the variability of stratospheric water vapor. The tropical tropopause temperature that could regulate the entry of tropospheric water vapor into the stratosphere is reported to have a cooling trend (Zhou et al., 2001; Seidel et al., 2001), indicating the opposite sign for what might be expected for a water vapor increase in the stratosphere. It has been reported, however, that the water vapor trend is not so evident if viewed from satellite and that there was a sudden drop in the stratospheric water concentration in the year 2001 (Randel et al., 2004). They also emphasized that the water vapor variations are consistent with the TTL temperature changes on a year-to-year basis. Fueglistaler and Haynes (2005) estimated the interannual variations of stratospheric water vapor by employing numbers of trajectory calculations and suggested that the stratospheric water vapor increase estimated from sonde observations could be an overestimate. In view of the great importance of water vapor in our climate system, however, much effort is required in trying to obtain a detailed description of stratospheric water vapor and in understanding the mechanisms of how it is controlled by examining the physical, chemical and radiative processes in the atmosphere.

The Lagrangian temperature history along trajectories has been a convenient measure for studying the dehydration of air parcels entering the stratosphere (e.g. Jackson et al., 2001; Jensen and Pfister, 2004; Fueglistaler et al., 2004, 2005; Fueglistaler and Haynes, 2005). However, there is little in situ water vapor data in the TTL over the western tropical Pacific to check the effectiveness of such calculations. In this paper, first results of the water vapor sonde observations in the western tropical Pacific, conducted as part of the Soundings of Ozone and Water in the Equatorial Region (SOWER) project in December 2003, are presented along with the trajectory calculations (Sect. 3). Problems in characterizing air parcels with the use of trajectories are investigated prior to it by examining typical examples (Sect. 2) intending to provide the

framework for interpreting the radiosonde data. A new approach, which we call the "water vapor match," is discussed briefly in addition to the efficiency of dehydration along the trajectories for the air parcels observed (Sect. 4).

2 Lagrangian characterization of air parcels

2.1 Horizontal scale of air parcels

In order to interpret the observed water vapor concentration by radiosondes in terms of dehydration along their excursion in the TTL, trajectory calculations provide useful information on the origin and the water vapor content of air parcels. In addition to the routes air parcels are supposed to follow, it is also important to consider whether the air parcel retains its identity or is modified due to irreversible mixing along the trajectories. A method that could help investigate such processes is the one that deals with a bundle of trajectories that are started simultaneously from multiple points surrounding the point of interest. Figure 1 shows two examples in which the air parcel is advected along the trajectories with its identity retained (top) and modified (bottom). In these calculations, four data points, each of which is distributed either 0.5° of longitude to the east or west or 0.5° of latitude to the north or south from Tarawa, Kiribati (1.35° N, 172.91° E) are considered. The region shown by connecting these 4 points with red lines has a horizontal extent of about 100 km and is hereafter referred to as the core of an air parcel. Additional 4 points connected by black lines surrounding the core, distributed 1.0° longitude/latitude apart from Tarawa, define the vicinity of the air parcel with a horizontal extent of about 200 km. This will include the maximum range of the radiosonde drift along a single flight. The outermost boundary is defined by 8 points either 2.0° of longitude/latitude separated from Tarawa or both 1.4° longitude and latitude distant from Tarawa, and represents a region of about 400 km of horizontal scale. This will correspond to the horizontal resolution of limb-viewing satellite instruments. The bundle of 16 forward trajectories thus defined are calculated on the 370 K isentropic surface

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based on the European Centre for Medium-Range Weather Forecasts (ECMWF) operational analysis. The trajectories are color-coded by the saturation mixing ratio (SMR) of water vapor estimated from the air temperature along the trajectories; cold (warm) colors correspond to low (high) temperature and low (high) SMR. The location and the shape of the air parcel are expressed by connecting the data points with a time interval of 24 h. The data points that define an air parcel initialized on 00:00 UT on 29 December 1998 over Tarawa (Fig. 1 top) are advected more or less in parallel to the west, turn to the north at around 120° E, passing the southern part of Vietnam 5 days later, and then turn to the east and move eastward over the sea south of Japan. The core of the air parcel retains the compact shape until it reaches Vietnam. But after it encounters the subtropical jet the shape is distorted due to the strong horizontal shear. During its movement to the west along the equator the SMR takes the value near 2 ppmv so that effective dehydration is expected to occur.

Another example shown at the bottom of Fig. 1 is the case in which the treatment of a single air parcel will be unacceptable. This illustrates a bundle of trajectories that are initialized at 00:00 UT on 15 December 1998 over Tarawa. The air parcel suffers from east-west elongation while it is advected to the south for 2 days, and then the distortion reaches such an extent that its core is hardly regarded as a single region. That is, the red-colored square that originally surrounded the core with a horizontal extent of about 100 km breaks up as the data points migrate away to reach thousands of kilometers of separation from each other. For visual clarity, the identification of the core region is omitted after two days since initialization in this case. This kind of circulation field must involve a horizontal divergence in the region where the air parcel is advected.

2.2 Effect of vertical shear

The investigation of the effect of vertical shear on tracking the air parcels will be more or less limited because the pressure levels available for describing the TTL are only coarsely located at 150, 100, and 70 hPa while the air parcels in TTL are often highly stratified. In addition, it is not easy to visualize the 3-dimensional structure in a single

diagram. Here an attempt will be made to describe the effect of vertical shear on trajectories by looking at those on four isentropes, 360, 365, 370 and 380 K on the longitude-height section.

Figure 2 is an example that shows the trajectories initialized at 00:00 UT on 29 December (top) and 15 December (bottom), 1998 over Tarawa previously examined in Fig. 1. Those shown on the latitude-longitude section are color-coded by the isentrope; 360 K in black, 365 K in red, 370 K in green, and 380 K in blue. On the longitude-height section below, trajectories are color-coded by the SMR of water vapor. The location of air parcels at 00:00 UT each day is indicated by dots on the trajectories. The horizontal projection of the four trajectories have a common feature that the movement is primarily controlled by the anticyclonic circulation. We can see that the outermost route is taken by those on 365 K isentrope (red), while the inner circles are drawn by those on the top (380 K) and the bottom (360 K). Thus the increase of potential temperature from 360 to 380 K displays appreciable difference in the trajectories.

The similar diagram for the case when the identity of the air parcel will be lost (Fig. 1 bottom) is shown in the bottom of Fig. 2. The feature of the southward drift is common to all four isentropes with the eastern routes taken by those on the lower two isentropes. Three days later, the difference between them becomes so large that those parcels on the lower 3 isentropes begin to move to the east while the one at the top travels to the west. Such a result might come from the fact that the location of divergence is different among altitudes and/or the stratospheric wind system is different from that in the troposphere. Thus the vertical wind shear plays an essential role in determining the fate of air parcels in such cases.

The characteristic features of a bundle of trajectories will thus help to estimate possible dehydration and irreversible mixing that have taken place for advecting air parcels. In the following section, in situ water vapor sonde data are examined with the use of backward trajectories corresponding to each sounding.

3 Analysis of the water vapor sonde data

In order to capture the air parcels being dehydrated during horizontal advection along the "cold trap," a series of radio soundings using water vapor sondes were conducted at Bandung, Indonesia (6.90° S, 107.60° E) and Tarawa, Kiribati in December 2003 as a part of the SOWER project. SOWER has been using the NOAA/CMDL frostpoint hygrometer (FPH) between 1998 and 2002 (Vömel et al., 2002) and the University of Colorado Cryogenic Frostpoint Hygrometer (CFH) since 2003 (Vömel et al., 2006[1]). It has also been flying the Snow White (SW) hygrometers (Fujiwara et al., 2003) since the year 2000, and has flown nearly 100 sondes. It is not a cryogenic frostpoint hygrometer such as NOAA's, but it utilizes a peltier-based thermoelectric circuit to make a frost on the mirror. Its advantage is that it works without the use of a liquid cryogen so that the preparation and operation for soundings are much easier than FPHs. It also helps to reduce the observational cost remarkably. The applicability of SW to the TTL water vapor observations has been studied by Fujiwara et al. (2003) and Vömel et al. (2003). Although it has a clear limitation in that the cooling efficiency is not high enough for stratospheric measurements, it will operate accurately down to the frost points at –75 to –80°C, that is, up to the middle TTL. In the December 2003 SOWER campaign, some SWs were launched simultaneously with CFHs to reduce observational uncertainties by collecting mutually independent water vapor data (Table 1). The launches in Bandung were scheduled in the morning to avoid possible shower, while in Tarawa they were conducted in the evening in order not to interfere with the routine meteorological service. Although the diurnal variation could possibly bring about systematic difference in the tropospheric features between the two stations, the TTL will be mostly free from tropospheric diurnal cycle as it is located above the altitude of main convective outflow.

Before presenting the sounding data, it is useful to take a look at the background me-

[1]Vömel, H., David, D., and Smith, K.: The University of Colorado Cryogenic Frostpoint Hygrometer (CFH): Instrumental details and observations, J. Geophys. Res., 111, submitted, 2006.

teorological condition during the campaign. Figure 3 shows the longitude-time sections of temperature (top) and zonal wind (second from top) on 100 hPa over the equator (area-weighted mean within ±6.25° latitudes) obtained from the ECMWF operational analysis for the month of December 2003. The crosses on the diagram mark the longitude and time corresponding to the sonde observations. The lower two panels are the same as the top except that the monthly mean values are subtracted for each longitude. We could see that the eastward propagating disturbances (lower panels) are superposed on the stationary wavenumber 1 structure (upper panels) with the temperature minimum in the western Pacific (dotted line on the third panel) and easterly (westerly) wind maximum around 120° E (120° W). The sonde observations in Bandung (the crosses on the left hand side) took place during the passage of this disturbance; the first sounding on 5 December may correspond to the westerly maximum (little temperature deviation) while the last one on 12 December may have taken place in the easterly maximum (little temperature deviation) after experiencing the temperature minimum in-between. Those in Tarawa (three marks on the right) took place appreciably earlier than the passage of this disturbance.

The time-height sections of several meteorological quantities as obtained from the ECMWF analysis are shown in Fig. 4, by interpolating the gridpoint values to the location at Bandung. The solid lines in the vertical direction indicate the time of sonde launches. The features such as the downward phase propagation, mutual phase relationship between temperature (left hand diagram in the lower row) and zonal wind (lower middle), and the lack of perturbations in the meridional wind component suggest that this disturbance is brought about by Kelvin waves. The TTL during the period from 8 December to 11 is experiencing the coldest phase of this event. On the other hand, no appreciable disturbances are observed over Tarawa (not shown).

According to the hypothesis of the dehydration pump by Kelvin waves (Fujiwara et al., 2001), the water vapor in the TTL during the campaign period will be saturated so that the intrusion of tropospheric moist air is inhibited by the closed "valve" shut off by Kelvin waves. Figure 5 shows the vertical distributions of the water vapor mixing ratio

(thick lines) given by CFH (solid) and SW (dashed), the SMR (thin lines) estimated from temperature by using Goff-Gratch equation (Goff and Gratch, 1946; Murray, 1967), and the ozone mixing ratios (dotted lines) observed at Bandung (top) and Tarawa (bottom). The water vapor profiles are shown only below 80 hPa pressure level as the data above it are not reliable. For the data at Bandung, those on 5 December (BD210) and 12 (BD217) are excluded because the CFH data are not available for the former and some instrumental problem is recognized for the latter. The CFH data above 110 hPa on 8 December (BD211) and the SW data on 11 December (BD216) are also omitted due to an instrumental problem. Although the water vapor data are rather noisy and some discrepancy between the CFH and the SW values is noticeable, we can see that the thick solid and dashed lines (observed water mixing ratio) almost overlap with thin solid lines (SMR) indicating that the upper troposphere above 150 hPa in Bandung is almost saturated. This is consistent with the expectation from the Kelvin wave effect mentioned above. In Tarawa, on the other hand, we can see a stepwise increase of the ozone mixing ratio above 120 hPa in day 10, probably reflecting the vertical propagation of small scale gravity waves. These perturbations occasionally decay in the lower TTL when irreversible mixing prevails. The gradual increase of the ozone mixing ratio above 120 hPa in the days 8 and 9 may be the result of such mixing following the preceding event. These kinds of waves will contribute to the net transport of ozone and dryness from the stratosphere to the upper troposphere (Hasebe et al., 2000). Actually we could see a dry layer spreading to reach 120 hPa level on the days 8 and 9. Although this height is close to the upper limit of the SW measurement range, the interpretation could be that the dry, ozone-rich stratospheric air is contributing to the dryness of the uppermost troposphere through the vertical transport by small scale waves and subsequent irreversible mixing.

Diabatic heating is generally small in the TTL (Gettelman et al., 2004a), although the heating rate strongly depends on the cirrus formation and the existence of underlying convective clouds (Hartmann et al., 2001). Therefore the day-to-day fluctuations of water vapor and ozone mixing ratios are better examined by changing the vertical

scale to the potential temperature. In this framework, any variability due to adiabatic vertical motion such as that brought about by Kelvin waves will disappear during the one-week campaign period in which the diabatic heating could be mostly neglected. Figure 6 shows the vertical distributions of temperature (left), ozone (center) and water vapor mixing ratios (right) given by CFH (solid lines) and SW (dashed). Top panels show the profiles taken at Bandung while those on the bottom are from Tarawa. Some trimming (as in Fig. 5) of water vapor data are applied due to instrumental limitations. The vertical distribution of temperature shows a specific structure exhibiting curvature changes at around the 345 K and 355 K isentropes. The atmospheric region between these isentropes corresponds to the lower TTL. The water vapor mixing ratio shows a rapid decrease with respect to height in this region. The profiles between 350 and 360 K isentropes in Bandung are in two groups; the one having higher water vapor values in the earlier days of the campaign (5 December (black) and 8 (red)) and the other with lower values during the later days (10 December (green) and 11 (blue)). This grouping does not apply, however, for the ozone mixing ratio; the profile for 5 December shows a decrease with respect to altitude while the others are almost constant in this height range with the values for 8 December a little higher. In Tarawa, the water vapor mixing ratio on 10 December is generally higher than the others below 360 K. These two profiles lie between the two groups in Bandung, although a dry layer noticeable near 360 K on 8 December and 9 discussed previously is an exception. The ozone profiles over Tarawa remain almost the same for three days.

The dry layer between the 350 and 355 K isentropes observed during the later period of the Bandung observations could be brought about by the "cold trap" dehydration as well as the instantaneous in situ effect. To examine if such differences could be interpreted by the level of coldness the air parcels have experienced during horizontal advection, the backward trajectories corresponding to the sonde observations are calculated on 355 K isentropic surface using the ECMWF operational analysis. The results are shown in Fig. 7 for 5 December and 8 at the top, for 10 December and 11 on the middle both for Bandung, and for all three launches from Tarawa at the bottom. The

trajectories for Bandung can be regarded as the case in which the air parcel retains its identity during advection. Trajectories corresponding to the relatively humid days of 5 December and 8 (top) show that the air parcels originated in midlatitudes have not been exposed to extremely low temperature (SMR of 7 to 10 ppmv), while those of the dry days of 10 December and 11 (middle) have spent a relatively long time advecting along the 5° S latitude circle and are processed by low temperatures (SMR of around 5 ppmv) just before the arrival at Bandung. Thus the temperature history that the air parcels have experienced could well be reflected in the observed water vapor mixing ratio. The trajectories for Tarawa, on the other hand, show a large scatter between the members in each bundle of trajectories, so that the identification of the core is abandoned for those earlier than four days before the sounding. Thus it suggests that the air parcels' identity was established just a few days earlier (Sect. 2.1). It is also seen that the temperature they have been exposed to is appreciably higher than that for the soundings at Bandung. The range of SMR during 24 h before arrival at Tarawa is about 16–23 ppmv for those parcels observed on the days 8 and 9 and about 20–33 ppmv for that of the day 10, again corresponding qualitatively to the observed difference in the water vapor mixing ratio around 350–355 K level (Fig. 7). All these facts are consistent with the interpretation that the difference in the observed water vapor mixing ratio reflects the temperature history of the observed air mass since its formation as well as the initial water vapor content. Although the current analysis deals only with dehydration in the lower TTL, and the dehydration that ultimately determines the stratospheric entry of water vapor might not necessarily be the same, the above results could be regarded as early observational evidence in support of the "cold trap" hypothesis by Holton and Gettelman (2001).

4 Discussion

The SOWER campaign in December 2003 took place during the passage of a Kelvin wave event over Bandung. However, the temperature anomalies (deviation from

monthly mean) at 100 hPa remain negative and the water vapor sonde data show near saturation in the lower TTL throughout the campaign period irrespective of the phase of Kelvin waves (Figs. 3 and 4). The lack of clear contrast in the relative humidity between the phases of the Kelvin wave event may imply that other processes such as horizontal advection dominate in controlling the saturation condition of the TTL in this case. Unfortunately, the observed water vapor data are too noisy to make a quantitative estimation of the dehydration at this altitude.

On the isentrope down to 350 to 360 K, however, two kinds of hygrometers, CFH and SW, show reasonable agreement. The observed water vapor mixing ratios on 355 K isentrope at Bandung are about 20 ppmv and 10 ppmv for 5–8 December and 10–11, respectively. The minimum SMR along the backward trajectories for the corresponding air parcels is 7–10 ppmv and 5 ppmv, respectively, thus indicating the existence of supersaturation and/or re-evaporation before sedimentation. These values suggest that the part of the water vapor retained in the air parcel after the "cold trap" dehydration is roughly twice as much as the minimum SMR. However, this estimate may include appreciable uncertainty due to the steep gradient of the water vapor mixing ratio in the 350 to 360 K potential temperature range and the temperature bias in the analysis field compared to observations. There could have been a moistening of air parcels during horizontal advection due to deep convection unresolved in the analysis field. The effect of wave-driven temperature perturbations unresolved in the analysis field (Jensen and Pfister, 2004) is another factor that makes an accurate estimate difficult. Information on the formation of TTL cirrus clouds, such as that given by lidars, will help reduce such uncertainties (Shibata et al., 2006[2]). Although mesoscale and microphysical processes might not necessarily be invoked in the Lagrangian estimation of annual mean and seasonal variation of stratospheric water vapor by trajectory calculations (Fueglistaler et al., 2005), it is necessary to accumulate this kind of ob-

[2]Shibata, T., Vömel, H., Hamdi, S., Kaloka, S., Hasebe, F., Fujiwara, M., and Shiotani, M.: Tropical cirrus clouds near cold point tropopause observed under supersaturated condition, J. Geophys. Res., 111, submitted, 2006.

servational evidences for justification of such simplified treatment and understanding of the interannual variability.

We have seen in Sect. 3 that the observed values of water vapor mixing ratio reflect the coldness the air parcels have experienced during advection. To make this argument more quantitative, knowledge of the atmospheric composition (water vapor and ozone) as well as the meteorological conditions along the trajectories is necessary. One approach to doing this would be repeated sonde observations of the same air mass following its motion. Such an idea was put forward and brought into practice for the study of ozone depletion in the stratospheric winter polar vortex. The method called the "match" technique (Rex et al., 1998) has contributed much in quantifying the influence of chemical loss and dynamical transport separately. In this method, the ozone destruction rate could be estimated with the use of the ozone continuity equation written in a Lagrangian form by quantifying the difference in the ozone concentration between the observations. A similar procedure could be applied to study the dehydration that processes the air parcels advecting in the TTL. However, we have to overcome several difficulties in applying this technique to studying dehydration in the TTL: (1) the accuracy of the meteorological field described by global analyses may not be as high in the tropics, (2) the dynamical field in the TTL is not as persistent as in the stratospheric polar vortex, (3) the predictability of the meteorological field may not be high enough to organize scheduled sonde launches, (4) upper air stations are distributed only sparsely in the western tropical Pacific and additional stations are not easy to implement, and (5) not many instruments exist that can measure water vapor in the TTL reliably and they are significantly more expensive than operational radiosondes.

In spite of these difficulties, we might expect that the horizontal scale of the air parcels, as measured by the temperature the air parcels are exposed to may be larger than that of the ozone destruction that is often affected by PSCs formed by small scale Lee waves. The uncertainty in judging the "match" condition may thus be not so strict. Instruments like the CFH that can measure water vapor in the TTL and the lower stratosphere are becoming more readily available and their cost may decrease (Vömel et al.,

2006[1]). Match pairs could be sought by taking the advantage of the difference in the trajectories among isentropes (Fig. 2), although the isentrope of "the best match" may not be the same as that originally intended at the time of sonde launch.

5 Concluding remarks

The accumulation of observational evidence on the distribution and variability of water vapor in the TTL is crucial for understanding the dehydration processes acting on air parcels entering the stratosphere. In situ observation by water vapor sondes is effective for this purpose due to their high vertical resolution and instrumental mobility. The Lagrangian characterization of air parcels suggests that the modification of air parcel properties could result from the branching-out and merging-in of nearby trajectories.

Water vapor sonde data taken in the western tropical Pacific during the December 2003 SOWER campaign have been analyzed to study the efficiency of "cold trap" dehydration by taking the advantage of four dual launches of SW and CFH (in addition to several single launches of SW alone). The day-to-day variations of the water vapor mixing ratio in the region between the 350 and 355 K isentropes can be interpreted on the basis of the origin of the air parcels and by the degree of coldness the air parcels are exposed to during horizontal advection. Although appreciable uncertainties remain, the observed air parcel may have retained the water vapor by roughly twice as much as the minimum saturation mixing ratio after its passage through the "cold trap." The "water vapor match" will better quantify the degree of dehydration observationally.

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Table 1. Summary of water vapor soundings during the December 2003 SOWER campaign.

Bandung Date	LT–7 Universal Time	Flight Num.	Hygrometer
5 Dec	0:27	BD210	SW
8	2:02	BD211/212	CFH, SW
10	1:19	BD213/214	CFH, SW
11	1:03	BD215/216	CFH, SW
12	1:49	BD217	CFH, SW
Tarawa	LT–12		
8 Dec	6:29	TR001	SW
9	6:12	TR002	SW
10	3:17	TR003	SW

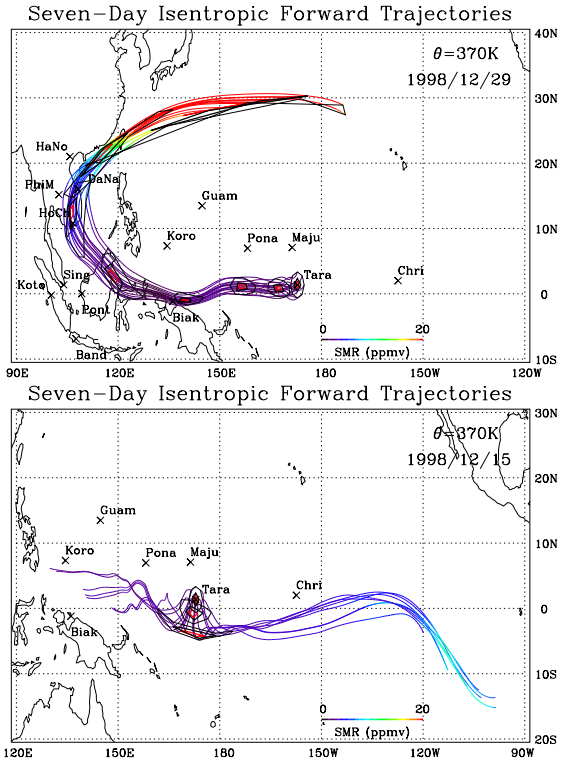

Fig. 1. Isentropic forward trajectories (370 K) initialized at 16 points surrounding Tarawa at 00:00 UT on December 29 (top) and 15 (bottom), 1998. The advective motion of the air parcel is visualized by identifying the location of its core (red square) together with its vicinity (black solid) for each interval of 24 h. See text for more explanation.

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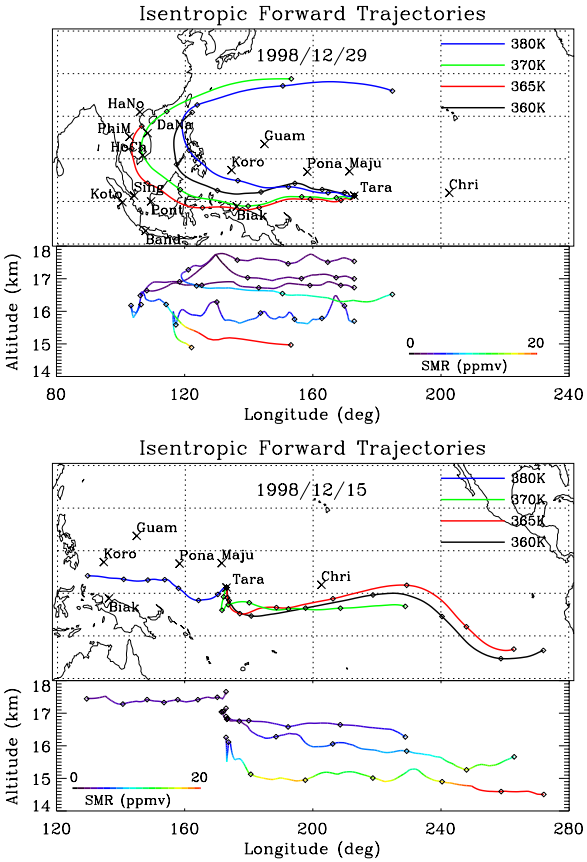

Fig. 2. Three-dimensional structure of air trajectories projected on the horizontal plane and the longitude-height section. The trajectories are initialized at 00:00 UT on 29 December (top) and 15, 1998 (bottom) over Tarawa on four different isentropes. Vertical projections are color coded by saturation mixing ratio. The advective motion is illustrated by dots for each interval of 24 h.

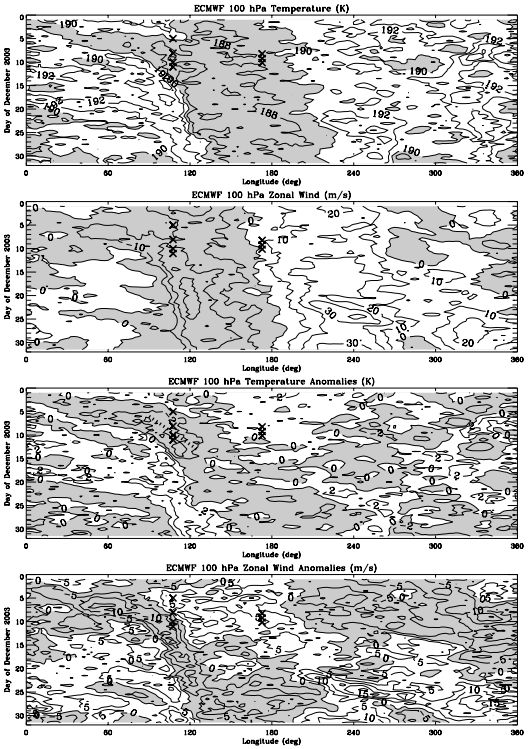

Fig. 3. Longitude-time sections of temperature and zonal wind (upper two panels) on 100 hPa over the equator (area-weighted mean within ±6.25° latitudes) being obtained from the ECMWF operational analysis for the month of December 2003. The lower two panels are the same as the top two except that the monthly mean values are removed for each longitude. The shading represents (from top to bottom panels) those area with temperature less than 190 K, easterly wind, negative anomalies of temperature and zonal wind. The crosses mark the longitude and time of the radiosonde observations. The dotted line on the third panel shows approximate location of temperature minimum associated with the Kelvin wave.

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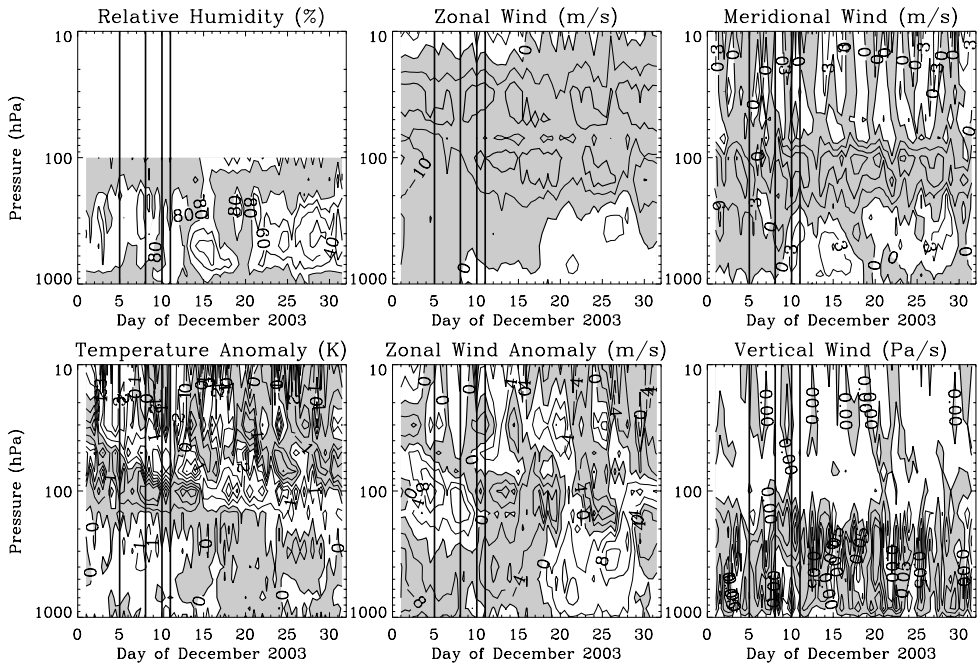

Fig. 4. Time-height sections of relative humidity (%), zonal and meridional wind ($m\,s^{-1}$) (upper row from left to right), temperature (K) and zonal wind ($m\,s^{-1}$) anomalies, and vertical wind ($Pa\,s^{-1}$) (lower row from left to right) for December 2003 interpolated to Bandung from gridpoint values of ECMWF analysis. Relative humidity above 100 hPa is omitted. The shading corresponds to those area with (top panels from left to right) relative humidity above 80%, easterly and northerly winds and (bottom from left to right) negative anomalies of temperature and zonal wind and upward wind component. Vertical lines mark the time of the radiosonde observations.

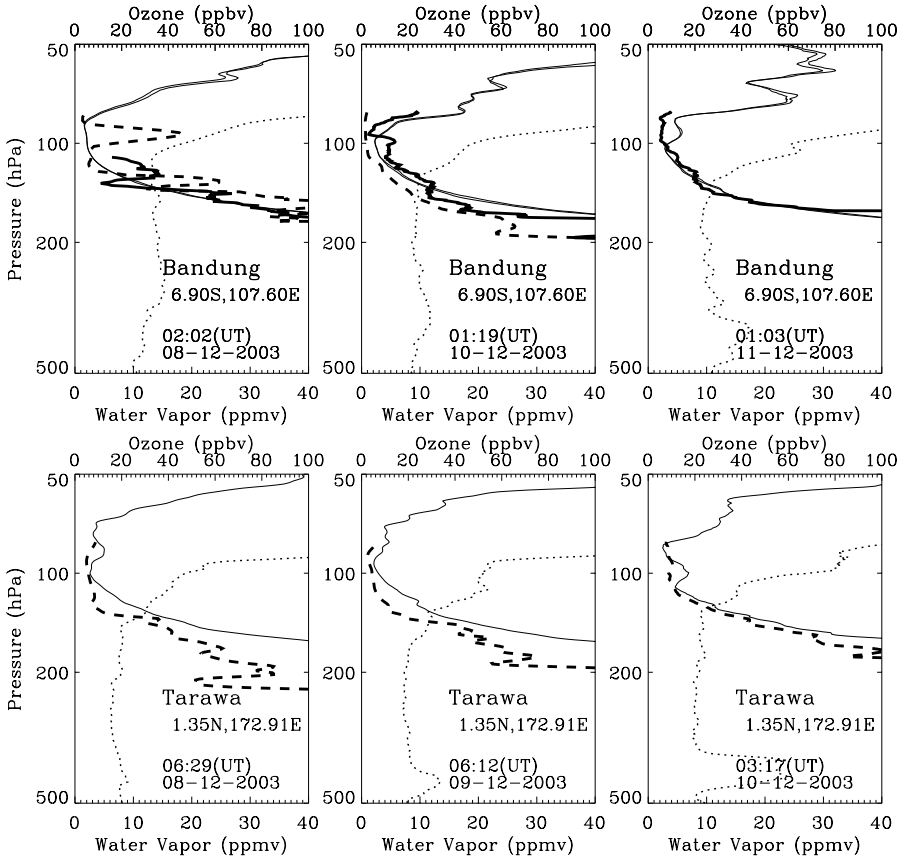

Fig. 5. Vertical distributions of water vapor mixing ratio (heavy lines) observed by CFH (solid) and SW (dashed) hygrometers, saturation mixing ratio (SMR) estimated from temperature (thin lines), and ozone mixing ratio (dotted lines) at Bandung (top) and Tarawa (bottom) in December 2003. Note that two SMR profiles appear on the diagram for Bandung due to the dual launch of radiosondes.

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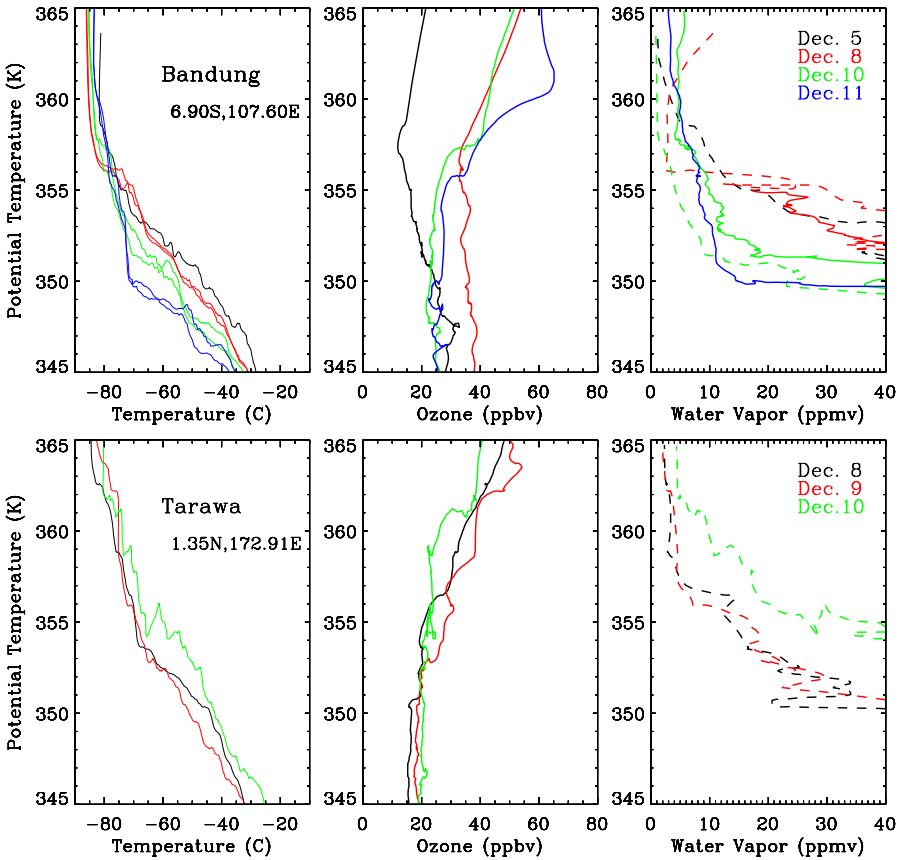

Fig. 6. Vertical distributions of temperature (left), ozone mixing ratio (center), and water vapor mixing ratio (right) obtained by water vapor sondes Snow White (dashed lines) and CFH (solid) in December 2003. Top panels show those from Bandung and the bottom are from Tarawa.

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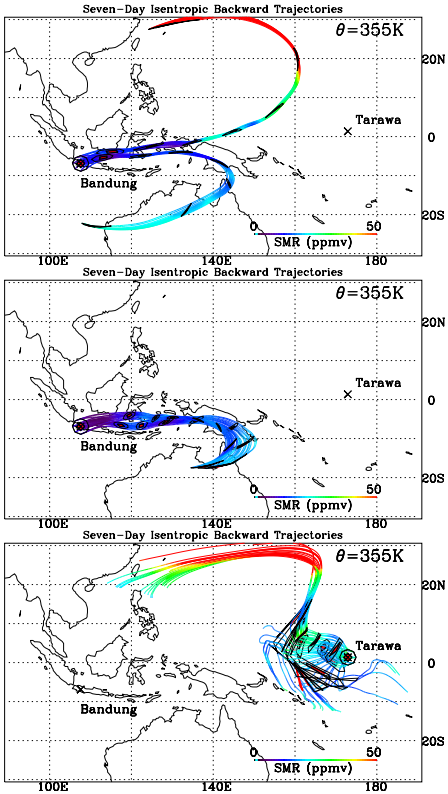

Fig. 7. A bundle of isentropic backward trajectories (355 K) corresponding to the soundings on 5 December and 8 (top), 10 December and 11 over Bandung (middle), and 8 December, 9 and 10 over Tarawa (bottom) color-coded by saturation mixing ratio of water vapor. See text for the details.