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Water vapor in the tropical lower stratosphere during the driest phase of the atmospheric "tape recorder"

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Abstract. Data from the Microwave Limb Sounder on the Upper Atmosphere Research Satellite are used to examine sequences of days during the driest phase of the atmospheric "tape recorder" signal in the lower stratosphere. It is found that the Indonesian region is the first place to become dry followed by the western Pacific and Panama. Eventually, dry air is found in a band along the equator with the most northerly extent over Indonesia and Panama. Two-dimensional trajectories show that winds cannot account for the spread of dry air from Indonesia and Panama across the entire tropics. The patterns are difficult to explain but may result from a combination of widespread ascent of air and the effects of deep convection.

1. Introduction

Water vapor in the stratosphere is a powerful tracer for atmospheric motions and its study has important implications for our understanding of the transport within the stratosphere and for stratosphere–troposphere exchange (STE). There are many unresolved problems concerning the exchange of air between the troposphere and stratosphere and in explaining the dryness of the lower stratosphere. This is partly because measurements of water vapor in the lower stratosphere are difficult to obtain. In situ measurements from balloon and aircraft are limited in their spatial and temporal coverage. Satellites have been increasingly exploited to provide better data coverage in this region. In this paper we present results from the Microwave Limb Sounder (MLS) on the Upper Atmosphere Research Satellite (UARS). MLS offers daily measurements of water vapor in the tropical lower stratosphere, and hence it allows us to construct daily maps and to follow synoptically the evolution of observed features.

The methods by which air gets into the stratosphere from the troposphere and becomes dehydrated to values typical of the stratosphere are still uncertain. *Brewer* [1949] found the observed mixing ratios to be lower than the minimum saturation mixing ratio at the local tropopause. To account for this dryness, he proposed a circulation whereby air enters the stratosphere from the troposphere in the tropics, drifts poleward, and descends in the extratropics. The tropical tropopause, which is high and cold, acts as a "cold trap" to "freeze dry" the air as it rises through.

The "extratropical pump" which acts nonlocally upon the tropical stratosphere [e.g. *Eliassen,* 1951; *Dickenson,* 1968] is responsible for large-scale ascent and mass transfer from the tropical upper troposphere to lower stratosphere. The extratropical pump drives tropical temperatures below radiative equilibrium and produces an annual cycle in tropical temperatures near the 100 hPa level, with lowest temperatures during northern winter and highest temperatures during northern summer. A consequence of the annual cycle in tropopause temperatures is that mixing ratios of air that has entered the stratosphere from the tropical troposphere should also have an annual cycle, in phase with that of tropopause temperature. Satellite observations [*Mote et al.,* 1995, 1996] indeed confirmed this. Layers of air retain this water vapor signal for up to 2 years, and this is known as the "tape recorder" effect [*Mote et al.*, 1996].

In the zonal mean, however, the lowest tropopause temperatures are not low enough to explain the lowest mixing ratios in the stratosphere [*Mote et al.,* 1996], a scenario which observations have long suggested and which led *Newell and Gould-Stewart* [1981] to define areas where the driest air could enter the stratosphere as being regions where the temperature was colder than average. They introduced the term "stratospheric fountain" and identified the most likely times and locations for the stratospheric fountain as being the western Pacific, northern Australia, Indonesia, and Malaysia from November to March, and over the Bay of Bengal and India during July and August, when the temperatures are 190.6 K at 100 hPa, cold enough to dry the air to 3 ppmv.

An alternative mechanism whereby tropical convective systems could cool the stratosphere was proposed by *Danielsen* [1982]. Some convective events might be strong enough to overshoot the tropopause, mixing tropospheric air with stratospheric air and leading to the formation of a large cirrus anvil in the stratosphere. The sedimentation of the ice crystals from the top of the anvil acts to dehydrate the region, and thus tropospheric air is introduced to the stratosphere but results in dehydration. On the basis of model simulations, *Potter and Holton* [1995] suggested that convectively generated buoyancy waves could also induce vertical parcel

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displacements which promote the formation of ice crystals in the lower stratosphere. Potter and Holton's mechanism, like Danielsen's, results in dehydration but does not require convection to penetrate the tropopause or reach the hygropause.

More recently, *Dessler* [1998] used radiosonde data from 1994 to 1997 to reexamine the stratosphere fountain hypothesis and found that the zonal and annual average mixing ratio of water vapor entering the stratosphere agreed well with the zonal and annual average saturation mixing ratio of the tropical tropopause region and therefore that the stratospheric fountain hypothesis was unnecessary. There remains some debate over this issue [*Vömel and Oltmans*, 1999; *Dessler*, 1999]. *Zhou et al.* [2001] analyzed radiosonde data over the years 1973-1978, a longer time period than that which *Dessler* [1998] used, and showed that although the stratospheric fountain might not have been necessary to account for stratospheric mixing ratios during the mid-1990s as Dessler found, it was necessary for most of the years before that. It has also been suggested that the word "fountain" is inappropriate since air has been shown to be sinking in the fountain region [*Sherwood*, 2000; *Gettleman et al.*, 2000]. The term "cold trap" might be more apt, with air mixing quasi-horizontally though the cold region [*Holton and Gettleman*, 2001].

Some of these theories lead more or less structure in the lower stratospheric water vapor field to be expected. The aims of this paper are twofold, to show that there is considerable structure to the water vapor field at 68 hPa and to exploit the daily coverage of MLS in the tropical region to follow synoptically the evolution of the observed features. We focus on the boreal winters of 1992 and 1993 when the driest air is found in the lower stratosphere. The appearance of the driest air has longitudinal preferences which could give some insight into the role of the stratospheric fountain in the dehydration of the lower stratosphere.

2. Data

The UARS satellite [*Reber*, 1993] is in an almost circular orbit at an altitude of 585 km and an inclination of 57° to the equator. It makes about 15 orbits a day. MLS makes a limb scan perpendicular to the UARS orbit path from a tangent height of 90 km to the surface. The measurements of limb radiance from one scan are used to deduce profiles of temperature and of the mixing ratio of various species. The profiles are retrieved on a fixed pressure grid. Version 104, which we use here, has six levels per pressure decade: approximately one level every 2.5-3 km. MLS provides a 3 km field of view in the vertical. Latitudinal coverage changes from between 80°N and 34°S to between 34°N and 80°S about every 36 days because of the satellite making a yaw maneuver. The tropical region is thus, barring occasional instrument problems, observed daily, and measurements are available from September 19, 1991 to April 22, 1993. The MLS instrument is described in more detail by *Barath et al.* [1993], and the measurement technique is described by *Waters* [1993].

Version 104 was a prototype version developed before the most recent version, version 5, and using a slightly different method. Version 104 is an improvement on previous versions [*Pumphrey*, 1999] and also has advantages over version 5. We have chosen not to use version 5 for this study because it appears to have more problems than version 104 in the lowest part of the stratosphere. A more detailed discussion of the use of version 104 and version 5 is given by *Pumphrey et al.* [2000].

For the reasons given by *Pumphrey et al.* [2000] we do not feel that the 100 hPa data can be believed sufficiently to investigate the spatial structure on a daily basis, although papers such as those by *Pumphrey et al.* [2000] and *Mote et al.* [1998a] have demonstrated that it shows some physically reasonable behavior. We therefore confine our study to the 68 hPa level. The averaging kernel for the 68 hPa level has a full width at half height of about 4.2 km, broader than the 3 km typical of the midstratosphere. The retrieved value at 68 hPa therefore depends to some extent on the true values at 100 and 46 hPa. It is important to keep in mind that the retrieved points should be interpreted as values at the breakpoints of a piecewise-linear representation of the vertical profile. We estimate, at this level, the precision of a single profile to be 0.3 ppmv and the accuracy to be 0.75 ppmv. MLS shows little bias against frost point hygrometer data at 68 hPa but shows a dry bias when compared with data from the HALogen Occultation Experiment (HALOE) and aircraft-mounted Lyman-alpha instruments [*Pumphrey*, 1999]. This should be borne in mind when considering the mixing ratios discussed in the remainder of this paper.

3. Evolution of Patterns at 68 hPa

The time-height sections which revealed the tape recorder effect in the paper by *Mote et al.* [1996] were based on the average water vapor mixing ratio in a tropical bin from 12°N to 12°S and hence gave no information about the horizontal distribution of water vapor and how it changes. At 68 hPa there remains some regional structure to the water vapor field. We describe this regional structure and its development over time.

The 68 hPa level at the launch date of UARS (September 8, 1991) was relatively moist, remaining so until around January 1992 [*Mote et al.*, 1996]. It then began to get drier, with the dry phase existing until June or July. This was followed by a moist tape signal which was present until January 1993. Sequences of days have been studied throughout the entire period for which there are data from the 183 GHz channel on MLS (September 1991 to April 1993), but we focus here on the later part of the transition from the moist to dry phases when the driest air appears at the 68 hPa level.

The longitude-time section in Figure 1 shows the evolution of this driest phase at 10°N from the beginning of January 1992. MLS footprints were interpolated onto points spaced every 5° in longitude at 10°N as described by *Clark et al.* [1998]. We shade contours in the range 2.8-3.4 ppmv

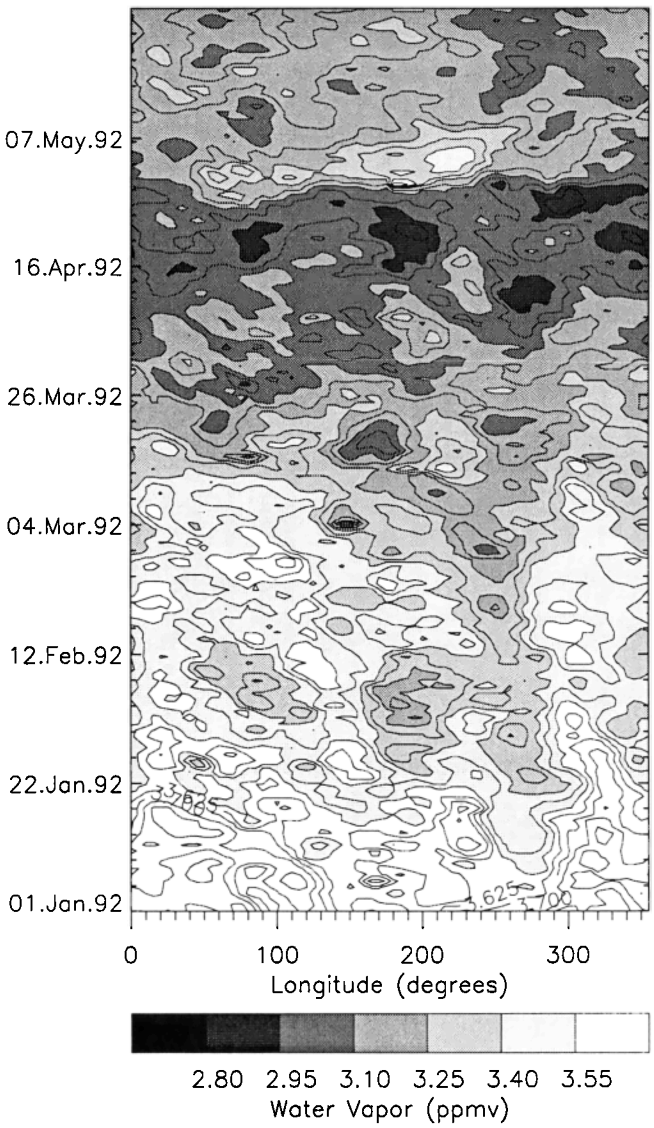

Figure 1. Longitude-time section at 10°N showing the driest phase of Northern Hemisphere winter 1992. The contour interval is 0.075 ppmv, and the shading corresponds to that in Figure 2.

to correspond with Figure 2. The slope of, for example, the 3.1-3.25 ppmv and 2.95-3.10 ppmv contours in early February 1992 suggests some eastward movement. There is frequent appearance of air with mixing ratios less than 2.95 ppmv over the Indonesian region around 130°E and in the eastern Pacific around 270°E. In late March there is a westward shift in the appearance of the dry air around Indonesia. The occurrence of these lowest mixing ratios is more sporadic and remains more localized compared with air of mixing ratio 2.95-3.10 ppmv which spreads across all longitudes by April. We look at the evolution of these features in more detail in section 3.1

3.1. Water Vapor Fields in 1992

Figure 2 shows 5-day averages of water vapor mixing ratio interpolated onto the 435 K isentropic surface, this surface corresponding most closely to 68 hPa during the time period of study. Maps were created using an exponential distance-weighted interpolation onto a grid of 2° in latitude by 10° in longitude. The use of 5-day averages, or pen-

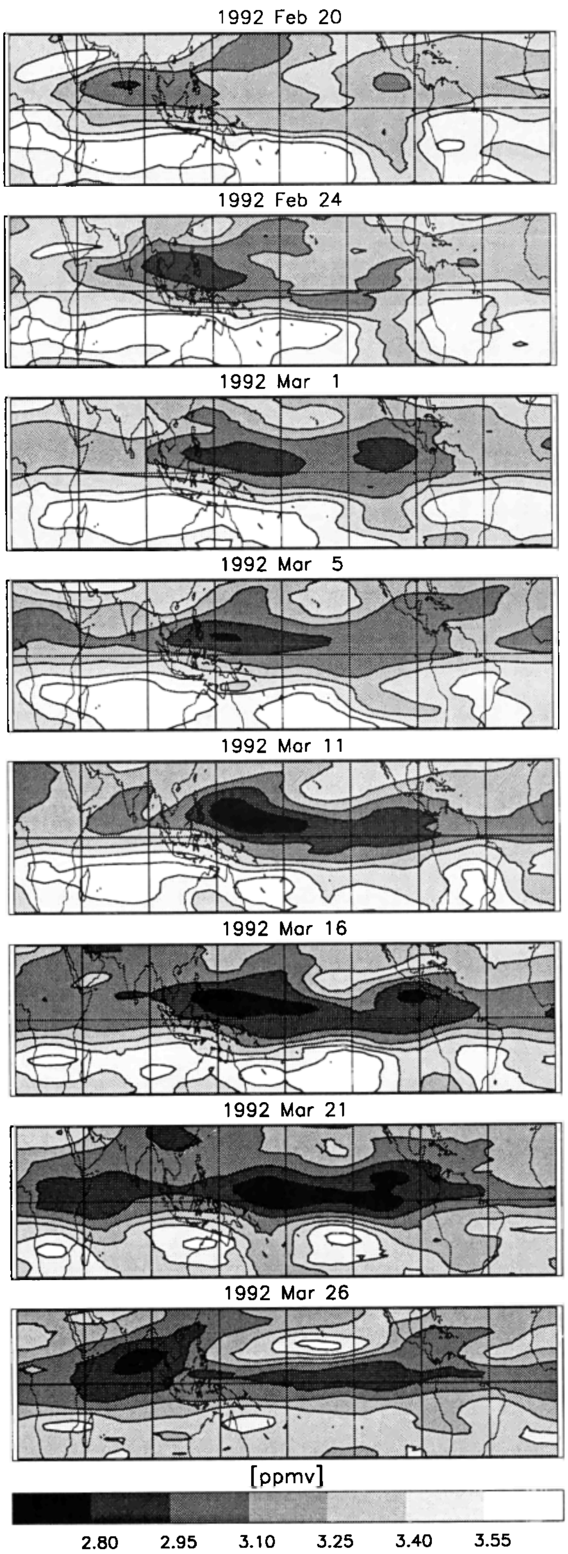

Figure 2. Sequence of 5-day averaged MLS water vapor (ppmv) on the 435 K isentropic surface showing the appearance of the driest air in 1992.

tads, removes some of the small-scale random features and improves the precision. The sequence begins on February 20, 1992, when air with a mixing ratio of 2.95–3.10 ppmv appeared in the Northern Hemisphere over Southeast Asia and Indonesia and over the eastern Pacific off the coast of Panama. These values spread rapidly, and seemingly eastward across the Pacific, and then westward toward Africa, eventually engulfing the entire latitude circle.

Drier air with mixing ratio 2.80–2.95 ppmv also appeared first over the tip of India followed by Indonesia, the coast of Panama on March 1, and Africa around the March 21. An eastward and southward spread of the pattern resulted in a dry band, confined to 15°N to 5°S along the equator, which remained until the middle of June. The appearance of the driest air (<2.8 ppmv) is more sporadic, seen first over the western Pacific on March 5. Looking at points later in the time series, air as dry as this has remained fairly localized, as the longitude-time section of Figure 1 suggests. Hence the lower stratosphere has dried to a background of 2.8–2.95 ppmv with occasional drier intrusions which have longitudinal preferences. Intriguingly, the new dry air is observed to appear in the regions identified by *Newell and Gould-Stewart* [1981] as the "stratospheric fountain".

The air is at most about 25% saturated, and the United Kingdom Meteorological Office (UKMO) temperatures would need to be at least 8 K colder to achieve saturation on the surface in question. This suggests that it is not drying in situ and is therefore likely to have come from below (not above since the gradient of mixing ratio is in the wrong sense), but some studies imply that that air might be sinking in these regions [*Sherwood*, 2000; *Gettleman et al.*, 2000]. Whether the vertical velocity introducing this dry air is an upward extension of the "fountain" or part of an unrelated large-scale uplift is beyond the scope of this study.

The longitudinal features are in broad agreement with January and April averages of water vapor at 70 hPa from the Stratospheric Aerosol and Gas Experiment (SAGE) II [*Rind et al.*, 1993], with dry areas being found above the convective zones of the western Pacific, Indonesia, Africa, and South America. Note that although SAGE II is used to form the a priori for MLS, the fact that it is a zonal mean implies that the longitudinal structure here is not imposed on SAGE II. *Jackson et al.* [1998] examined seasonal, multiyear averages from HALOE at 128 hPa, 100 hPa, and 83 hPa. They noted that at 100 hPa the lowest water vapor values (around 2.4 ppmv) appeared in the Northern Hemisphere to the west of the Indonesian source region and that the dry air then spread southward. Both SAGE II and HALOE, however, have a much more limited spatial and temporal resolution in the tropics than MLS and are not able to provide the daily coverage of the tropical region that we present here.

Figure 3 shows the difference of the 5-day averaged MLS fields from the averaged fields from 10 days before. Much of the tropical region had values less than zero, which is indicative of a general drying. Changes of between -0.25 and -0.15 ppmv took place in the Indonesian region on February 24, and the area moved eastward across the Pacific Ocean in the following pentad. At the same time the Indian Ocean region saw a slight increase in moisture which was followed by a reduction in moisture of 0.25 ppmv on March 16 and 21. Southern Africa and South America were also subject to a drying of less than -0.25 ppmv on March 21. On March

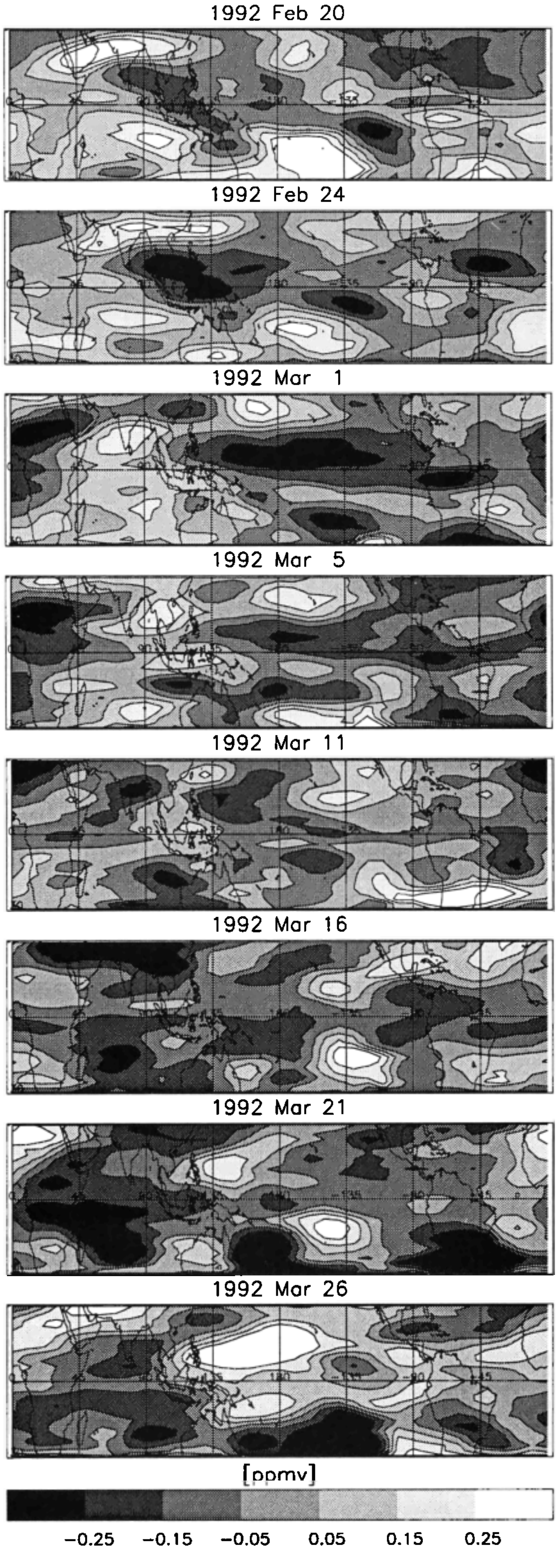

Figure 3. Difference between the MLS water vapor field (ppmv) and that of 10 days earlier. This figure therefore illustrates the change in the observed field over 10 days.

26 we notice an area of moistening located over the northern Pacific in the region of Hawaii.

3.2. Horizontal Advection for 1992

We investigate the extent to which these dry and moist changes in Figure 3 are produced by horizontal advection and how, in particular, the dry area observed by MLS over India on February 20 in Figure 2 spreads under the influence of the zonal winds. On inspecting Figure 2, there is an impression that dry air arrives on the February 20 near India and spreads eastward. We investigate whether the apparent eastward movement can be accounted for by the horizontal winds alone or whether vertical motion, mixing, or nonconservation of water vapor need to be invoked to explain the evolution. We address this using horizontal trajectories derived from the UKMO horizontal wind fields.

The UKMO meteorological analyses were produced using the method of data assimilation to complement the UARS project [*Swinbank and O'Neill,* 1994a]. The use of data assimilation allows information to be inferred about aspects of the atmospheric circulation which are under-represented by observations. In the tropics the wind fields are heavily dependent on the model because of the lack of observations there, but *Swinbank and O'Neill* [1994b] showed that the analyzed winds agree well with the available radiosonde observations in the lower stratosphere. The analyses were also shown to contain a realistic quasi-biennial oscillation (QBO) in the middle and upper stratosphere and semiannual oscillation in the upper stratosphere and lower mesosphere.

Coy and Swinbank [1997] compared UKMO winds with those assimilated from the Goddard Space Flight Center (GSFC) for February 1992 (the time period of study here). They found that zonal winds in the lower stratosphere compared well, with those from GSFC being greater than those from UKMO by 2 ms^{-1} at about 100 hPa and those from UKMO being greater than those from GSFC by 2 ms^{-1} at around the 68 hPa level. They also compared local differences between the two data sets by calculating the difference between the two wind components at each grid point for each day during the month. At about 68 hPa in the tropics, GSFC winds were 3-5 ms^{-1} greater than those UKMO.

In addition, an unpublished study by one of the authors (A. Billingham) of the differences between the UKMO assimilated winds and the Singapore radiosondes at 68 hPa gives some confidence that the trajectories can be adequately computed. The December-January-February period was studied for 1997, 1998, and 1999 (the times of our available radiosonde data), and the results from the comparison of the 1997 period are offered as representative of the study's findings. The mean zonal wind speed at the Singapore radiosonde station was 6.7 ms^{-1}, and the mean meridional windspeed was 2.5 ms^{-1}. The mean difference between the radiosonde and UKMO wind components (radiosonde minus UKMO) was 1.3 ms^{-1} for the zonal wind speed and -0.4 ms^{-1} for the meridional wind speed, with the corresponding RMS of the random differences being 3.4 ms^{-1} and 2.9 ms^{-1}, respectively. The correlation between zonal wind components was 0.86.

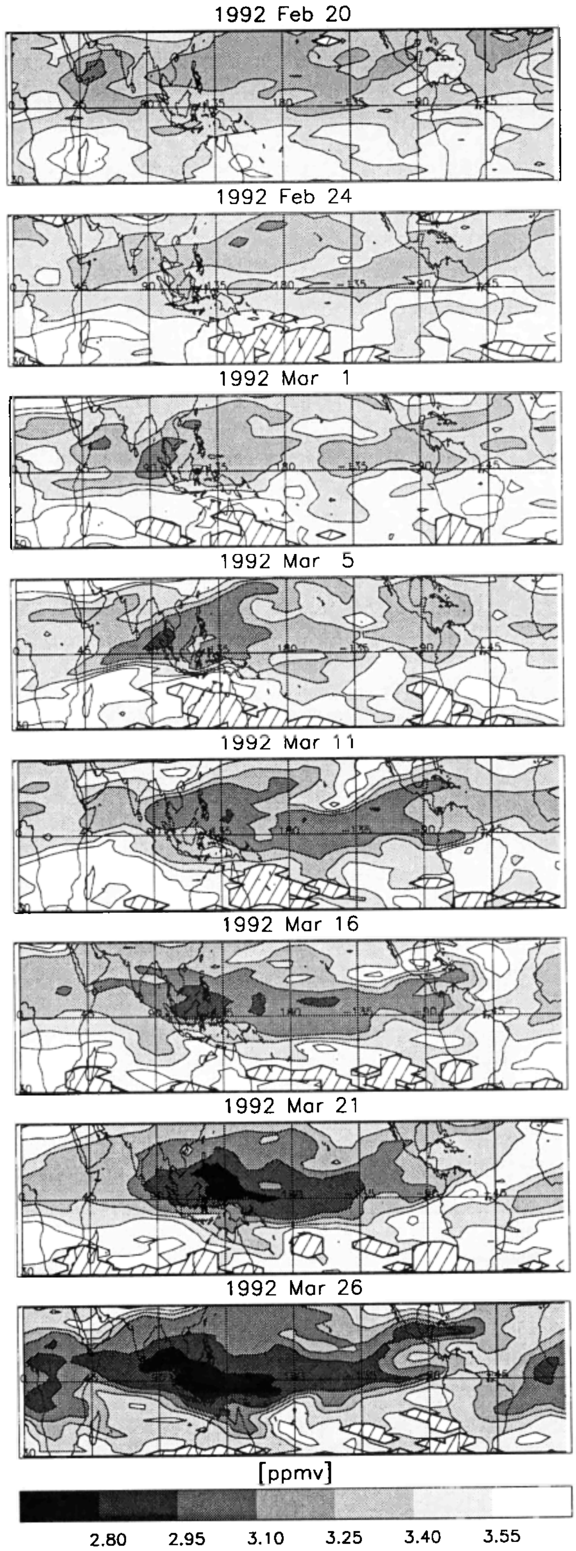

Figure 4. Ten-day advection of MLS water vapor (ppmv) for 1992 on the 435 K isentropic surface using UKMO winds.

If the bias between the radiosonde and the analyzed wind at Singapore is added to the UKMO wind field everywhere, it results in a displacement of the advected field by about 1000 km in the zonal direction (1000 km being within the 2600 km resolution of MLS at the equator) and 300 km in the meridional direction over a 10-day advection. Similarly, perturbing the winds by an amount comparable to the difference between the two data sets does not make a significant difference in the resulting fields. We therefore believe the UKMO analyses to be sufficient to support the qualitative statements that we make below.

Figure 4 shows the water vapor field after it has been advected for 10 days on the 435 K isentropic surface with UKMO horizontal wind fields using a fourth order Runge-Kutta scheme. The UKMO winds are updated once every 24 hours. Following *Waugh and Plumb* [1994], we estimate that reasonably accurate trajectories can be calculated with daily wind fields and that significant differences develop only when the wind field is updated after more than 1 day has elapsed.

For each of the 5 days that comprise the pentad, the MLS footprints from 10 days before were interpolated onto parcel positions which, when advected forward with the UKMO winds, would leave the parcels on the same regular grid used in Figure 2. The advection is performed for 10 days, and the resulting fields are averaged over 5 days. Figure 2 and Figure 4 can thus be compared. The production of filaments leads to details which cannot be resolved with MLS, and we therefore smooth the advected field slightly to give a comparable view to that from MLS.

The hatched areas in Figure 4 are areas where no water vapor value has been assigned to the parcel because the parcel began its journey outside the viewing range of MLS on 3 or more of the days which comprise the 5-day average. When MLS is looking north, the hatched areas occur in the Southern Hemisphere, and when MLS is looking south, they occur in the Northern Hemisphere. The hatched areas in Figure 4 tend to coincide with some of the highest mixing ratios, replacing regions of mixing ratios greater than 3.70 ppmv in Figure 2. This suggests that the highest mixing ratios result from transport into tropics through the barriers of the "tropical pipe" [*Plumb,* 1996]. The spreading of the tape recorder signal out of the tropics to midlatitudes was seen in zonal means of MLS data at 68 hPa by *Pumphrey et al.* [2000].

Values of less than 3.10 ppmv are widespread throughout the sequence of maps from MLS (Figure 2), whereas in the sequence of advected maps (Figure 4), mixing ratios of similar value are not prominent until March 5, some 20 days into the sequence. Similarly, mixing ratios in the observed field of less than 2.95 ppmv are not obvious in the advected field until March 16, some 25 days after they first appeared in the observed field. We note that the advected fields (Figure 4) do not preserve the maxima and minima of the initial fields (Figure 2) although the process of horizontal advection should preserve them. We have investigated the apparent discrepancy in detail, and there are two processes which contribute to the apparent nonconservation. Firstly, the 5-day time average is a contributor, and secondly, although values are correctly preserved by individual trajectories, the production of filaments followed by the spatial averaging implicit in the mapping technique alters the mapped maxima and minima.

On March 1, in the observed water vapor field (Figure 2), mixing ratios of less than 3.10 ppmv are found across the breadth of the Pacific Ocean. Air with these mixing ratios is seen only in the Indian Ocean and Bay of Bengal region in the advected field (Figure 4). On March 5, air with a mixing ratio of 2.95–3.10 ppmv has spread right along the equator from its more localized position on February 24, 10 days earlier. The corresponding advected field (March 5) shows that the 2.95–3.10 ppmv area has remained confined to the Indonesian region (with some advection along the coast of China). Similarly, the air with mixing ratios of 2.80–2.95 ppmv which covers the Pacific in the observed field on March 11, has spread all round the world by March 21st. The advected field for March 21 shows the 2.80–2.95 contour interval to still be centered over the Pacific Ocean.

We see that horizontal advection by UKMO winds cannot account for the evolution of the patterns in the water vapor field observed at 435 K. Thus they do not support the impression that air spreads solely from Indonesia and the western Pacific, which were regions identified by *Newell and Gould-Stewart* [1981] as the most likely places where air enters the stratosphere. If not by advection, it could be inferred that the appearance of dry air outside these regions has originated at a different altitude. The MLS data suggest that the dry air probably comes from below 435 K since the mixing ratios decrease with height. This can be inferred from the phase of the tape recorder and seen on inspecting fields of MLS water vapor at 46 hPa.

If air rises through the tropopause continually and in most places, as implied by the tape recorder [*Mote et al.,* 1996], the spatial pattern we observe would be the consequence of a general slow ascent of a zonally asymmetrical pattern imposed lower in the atmosphere, probably by convection. It would then be surprising for that spatial variation to have remained up to 435 K. Taking the tropical tropopause to be around 380 K [*Holton et al.,* 1995], we would have expected rapid zonal motion to have smeared out any pattern over the 60 days or so [*Mote et al.,* 1998b] that it would take for the air to ascend to the 435 K surface studied and, as a result, for the 68 hPa surface to have little longitudinal variation and a mostly zonal structure.

Since localized patterns are observed, we are led toward two possibilities. The first is that because of the limited vertical resolution of MLS and the thickness of the weighting function, the patterns seen at 68 hPa might be the result of contamination by features which actually occur lower in the atmosphere. The second is that the patterns are indeed imposed close to the 68 hPa level by convection or by some other mechanism.

To consider the first issue, some indication of whether the patterns we observe at 68 hPa are genuinely near that surface, or are retrieval artifacts imposed from a different level, can be obtained from the averaging kernels in Figure 5. The retrieved profile is piecewise linear. The averaging kernel shows the weight with which variations in the true profile (in this case represented by a similar piecewise-linear profile) contribute to variations at the nodes (join points). Ide-

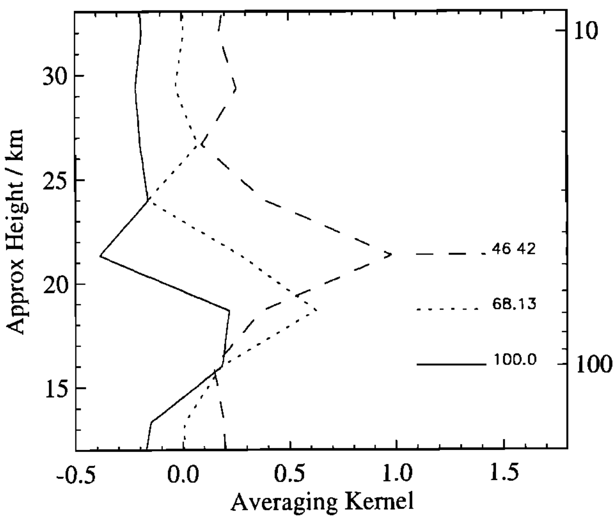

Figure 5. Averaging kernels for 68 hPa and the two levels on either side. The 100 hPa and 46 hPa kernels have been offset by -0.2 and +0.2, respectively, to make the figure clearer.

ally, the averaging kernels for a given node would be zero at all the other node points. Figure 5 shows that the kernel for 46 hPa approximates that fairly well, while those for 68 hPa and 100 hPa are less ideal. In particular, the 68 hPa kernel has the value 0.63 at 68 hPa and 0.18 at 100 hPa. Thus a perturbation of about 2 ppmv in the true profile at 100 hPa produces about a 0.6 ppmv perturbation in the retrieved profile at 68 hPa.

To further demonstrate these effects, we apply a sinusoidal wave perturbation in the longitudinal direction at 100 hPa to an otherwise unperturbed atmosphere. Figure 6 shows how this appears in the retrieved values at 100 and 68 hPa. The effect of the perturbation can be seen at both levels, but with amplitude diminished by values consistent with the averaging kernels. While it is true that the retrieved patterns at 68 hPa could be induced by a larger perturbation at 100 hPa, in that case we would expect to see signs of the same pattern in the retrieved values for the 100 hPa level. The 100 hPa fields show some correlation with 68 hPa, but variations are of smaller amplitude. The contribution from lower levels is small, and we therefore infer that the patterns are imposed close to 68 hPa or, at least, well above the 380 K or 90 hPa levels traditionally used as the demarcation between the troposphere and stratosphere.

There is much discussion as to the height to which tropical convection can reach, but generally, it is not considered to be high enough or occur with sufficient frequency, such that it could impose patterns on the 68 hPa surface, as we have seen here. *Danielsen* [1982] suggested that tropical convective systems could lead to dehydration in the stratosphere, prompting in situ measurement campaigns to look for evidence of deep convection in the tropics. The Stratosphere-Troposphere Exchange Project, Tropical Experiment (STEP Tropical) [*Russell et al.,* 1993], for example, investigated stratosphere-troposphere exchange and the dehydration process in the western Pacific and northern Australia during the monsoon in January–February of 1987. *Pfister et al.* [1993] showed that monsoon tropical cyclones mixed tropospheric air directly into the stratosphere at heights of around 18 km by means of detrainment from small and numerous overshooting turrets.

Convection that reaches these altitudes is thought to be infrequent, with some studies suggesting that tropical convection rarely reaches above the tropopause [e.g., *Highwood and Hoskins,* 1998] and with little extent beyond about 14 km [*Folkins et al.,* 1999]. Although cirrus cloud formation is frequently observed at 68 hPa, as *Wang et al.* [1996] and *Mergenthaler et al.* [1999] have seen with satellite data, it is unclear whether this cirrus forms as a result of overshooting convection or as the result of slow wave-driven ascent. A. Gettleman et al. (Distribution and influence of convection in the tropical tropopause region, submitted to Journal of Geophysical Research, 2000) found that in convective regions, clouds penetrate the tropopause only 2% of the time and cannot supply sufficient mass to the tropical lower stratosphere. *Sherwood and Dessler* [2001], on the other hand, suggest that the layer between the typical detrainment height (150 hPa) and 100 hPa could be replaced by dry air from overshooting convection by the time air had crossed through it by slow ascent, and they state that this argument also holds for the lower stratosphere. Much debate surrounds the issue of overshooting convection and its relative contribution to the water vapor content of the lower stratosphere. The interpretation of the features in the water vapor field shown here rests crucially upon this debate.

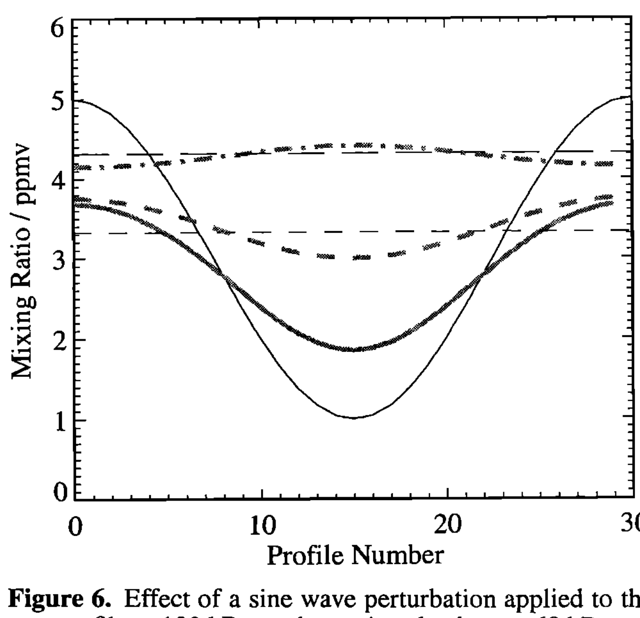

Figure 6. Effect of a sine wave perturbation applied to the true profile at 100 hPa on the retrieved values at 68 hPa and 46 hPa. The thin lines show the true state at 100 hPa (thin, solid line), 68 hPa (thin, short-dashed line) and 46 hPa (thin, long-dashed line). The 100 hPa line is therefore the applied perturbation, and the 68 and 46 hPa lines are constant. The thick lines show the retrieved value after the perturbation is applied, at 100 hPa (thick, solid line), 68 hPa (thick, dashed line), and 46 hPa (thick, dash-dotted line).

The patterns that we see at 68 hPa could well be the result of both overshooting convection and slow ascent and would therefore perhaps be better interpreted within the "tropical tropopause layer" (TTL) theory as suggested by *Sherwood and Dessler* [2000]. They propose that air detrains at various levels throughout the TTL, defined as the area between about 14 km, where most convection detrains, and 70 hPa, the highest level that convection can reach. Air on any given isentropic surface will have a range of "ages", with air that is further from convection, having been in the TTL for longer, being older and therefore moister than air that has recently detrained. The patterns we describe are found just below the top of their proposed TTL. Figures 1 and 2 showed that air in the two main convective regions, the western Pacific and Panama, was drier and therefore probably younger, having arrived on the 435 K surface first, just 10 days or so before air which detrained elsewhere within the TTL. To produce the driest areas over the western Pacific and South America, which are fairly consistent in Figure 1, however, air in the convective regions would have to be constantly replenished by new dry air from convective detrainment at a level close to 68 hPa, and it is unclear as to whether this would be possible within this theory. Air outside the convective regions must have taken longer to have come up from below, having detrained from tropical convection near to the bottom of, or having risen slowly from the bottom of, the TTL. It may have taken a longer and more quasi-horizontal route [*Holton and Gettleman,* 2001], passing through the "cold trap" region at some point on its ascent.

3.3. Water Vapor Fields in 1993

The Northern Hemisphere winter period for 1993 is shown in Figure 7. Because of a couple of days on which MLS was not operating, we begin the 1993 sequence on February 23, 3 days later than in 1992. Although mixing ratios of less than 2.95 ppmv were prominent at 68 hPa between February and August 1992, values of 2.95-3.10 ppmv are much less widespread in 1993. The moist areas in the southern subtropics are similarly located in 1992 and 1993 over Africa, South America, and the South Pacific. The dry features, however, show little eastward translation, and there is no zonal structure, which was so striking in 1992. The driest areas again occur first over Indonesia and the western Pacific, followed by Central America.

The advected fields are shown in Figure 8. As we saw in 1992, the advected field is not as dry as the observed field. On March 15, for example, we see that the observed field shows mixing ratios of 2.95-3.10 ppmv in comparison to the advected field, with lowest values of 3.10-3.25 ppmv. Moreover, the advected field on March 15 looks very similar to the observed field on March 5, 10 days previously, revealing that there has been very little change by advection over the 10 days.

Although the maximum mixing ratios in 1992 and 1993 are similar, the minimum mixing ratios in 1992 are much lower than those in 1993. Some interannual differences in the degree of dryness were seen in the zonal mean water vapor from MLS at 100 hPa and 68 hPa [*Pumphrey et al.,* 2000]. The reasons for this interannual difference may be related to one or more possible causes. The eruption of Mount Pinatubo in June 1991 increased the aerosol loading of the lower stratosphere and led to a warming [*Labitzke and McCormick,* 1992; *Angell,* 1993]. A consequent increase

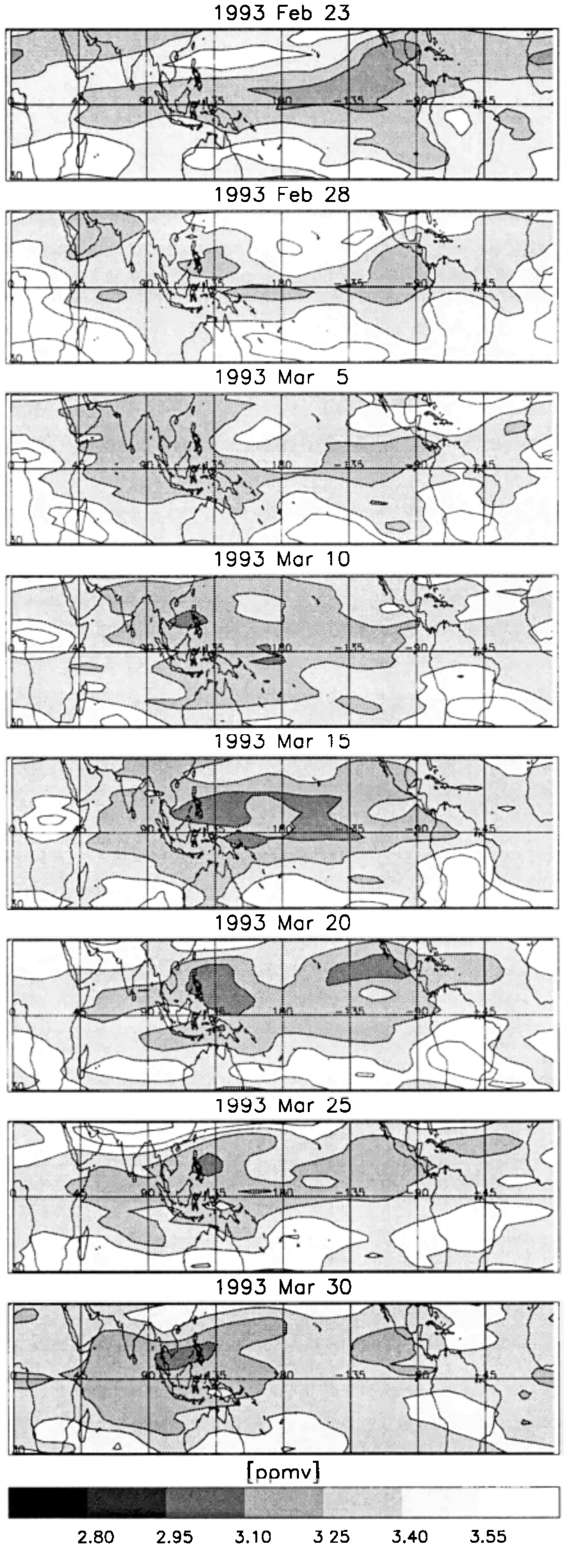

Figure 7. Sequence of 5-day averaged MLS water vapor (ppmv) on the 435 K isentropic surface showing the appearance of the driest air in 1993.

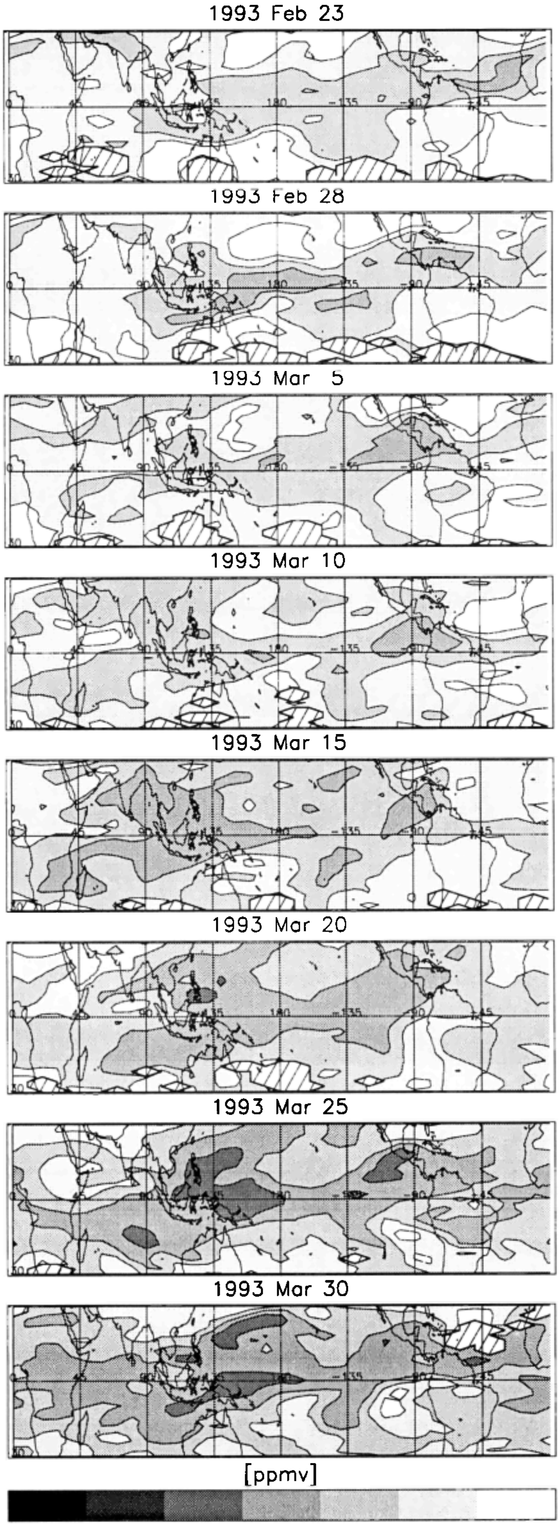

Figure 8. Ten-day advection of MLS water vapor (ppmv) for 1993 on the 435 K isentropic surface using UKMO winds.

in stratosphere-troposphere exchange may have resulted in increases of methane and of water vapor through methane oxidation [*Schauffler and Daniel,* 1994], but an increase of water vapor has not been observed by MLS [*Elson et al.,* 1996; *Evans et al.,* 1998]. *Mergenthaler et al.* [1999] noted that cirrus clouds were more prevalent during the DJF period of 1992-1993 than in the same period in 1991-1992 and suggested that because of the presence of the aerosol veil following the eruption of Mount Pinatubo, less solar radiation would penetrate to the tropics and less convective activity would reach the 68 hPa level, resulting in less dehydration in 1991-1992. Hence the effect of the eruption of Mount Pinatubo would be likely to increase water vapor or reduce drying by dehydration, and so we might expect 1991-1992 to be wetter than 1992-1993, opposite to what is observed.

It is also possible that the interannual dryness is related to the phase of the quasi-biennial oscillation (QBO). When the QBO is in its easterly phase, as it is during boreal winter 1991-1992, lower equatorial temperatures and greater ascent rates are implied [*Carr et al.,* 1995]. During the westerly phase of the QBO, temperatures are higher, and ascent rates are lower. This was the case in northern winter 1992-1993. The easterly phase of the QBO could therefore account for the lower mixing ratios in 1991-1992 if temperatures were lower. The easterly phase of the QBO may also encourage deep convection [*Collimore et al.,* 1998] to overshoot the tropopause and hence lead to a drier lower stratosphere. There is, however, much uncertainty over the frequency of convective overshooting events.

The phase of the El Niño and Southern Oscillation (ENSO) could also have an effect on the dryness in the lower stratosphere because of changes to the temperature and height of the tropopause [e.g., *Reid and Gage,* 1985; *Gage and Reid,* 1987]. These interannual variations are not apparent in the zonal mean but are evident as an east-west shift in the longitude of the height and temperature perturbations [*Gage and Reid,* 1987; *Randel et al.,* 2000]. ENSO affects also the height, location, and frequency of deep convection and cirrus formation [e.g., *Wang et al.,* 1996; *Mergenthaler et al.,* 1999]. The El Niño conditions of DJF 1991-1992 [*Trenberth,* 1997] and even the strong Madden-Julian oscillation at the same time [*Clark et al.,* 1998; *Mote et al.,* 2000] may well have encouraged convection, colder tropopause temperatures, and dehydration and altered the longitudinal structure of the patterns of water vapor observed. *Gettleman et al.* [2001] noted the spatial difference of the minima in water vapor observed by HALOE at 82 hPa. During El Niño they found that the minimum in water vapor extended from the maritime continent to the coast of South America and that during La Niña conditions it was located north of the equator in the western Pacific extending into the Indian Ocean. They found that at this level the minimum mixing ratios were higher during El Niño conditions, contrary to what we saw with 1992 and 1993 from MLS at 68 hPa. However, with only 2 years worth of data, is impossible to ascribe these interannual differences to either ENSO or the QBO since they operate on a longer timescale than that for which we have data available. The reason for such a marked difference in dryness between the Northern Hemisphere winters of 1991-1992 and 1992-1993 is clearly something which requires further investigation.

4. Summary

During the drying phase of both 1992 and 1993 at 68 hPa, MLS shows that the first region to become drier is the Indonesian fountain region followed by the western Pacific, Panama, and the western coast of North America. Moister areas continue to remain over the Southern Hemisphere continents and the southern Pacific Ocean. In 1992, dry air is eventually found in a band along the equator with the most northerly extent over Panama and Indonesia, but there is no such zonal structure evident in 1993.

Some differences in observed dryness exist from year to year. Winter 1991-1992 was much drier than 1992-1993, and this could be attributed to the easterly phase of the QBO in 1991-1992, lowering the temperature of the tropopause and encouraging deep convection. ENSO events or the eruption of Mount Pinatubo may also have played a part in causing 1992-1993 to be moister than 1991-1992, but with the limited length of the available data set it is not possible to assess the importance of these effects.

Two-dimensional trajectories from UKMO analyses show that air of high water vapor mixing ratio is advected into the tropical region from outside the 30°N to 30°S band. The trajectories reveal that low-mixing-ratio air is not spread from a source region over Indonesia across the western Pacific to account for the dryness over Panama observed some 10 days later. Horizontal transport alone cannot account for the observed patterns. We infer that there must be vertical transport of water vapor in numerous places across the tropics to account for the observed pattern. We suggest that the importance of the stratospheric fountain region is not that all or most of the air enters the stratosphere there, but more that it is the first place that sees low mixing ratios.

The fact that MLS shows such patterns in the water vapor field at 68 hPa is somewhat surprising and is difficult to explain. If the air rose gradually under the action of the extratropical pump, we would have expected rapid horizontal motion to have smeared out any pattern, leading to a mostly zonal structure to be observed at 68 hPa. We suggest that the pattern in the water vapor field is imposed higher than the traditional level of 380 K used to indicate the tropopause and well above the 14 km level at which most convection is considered to detrain. We suggest that the patterns could be interpreted as resulting from the action of both slow ascent and the action of vigorous convection creating a mixed "tropical tropopause layer" like that proposed by *Sherwood and Dessler* [2000]. The results must be interpreted with some caution, however, because of the limited vertical resolution of MLS. We await the launch of the Earth Observing System Microwave Limb Sounder (EOS MLS), whose improved vertical resolution will enable such issues to be addressed. We also recognize that threedimensional trajectories would make an important contribution to the understanding of the transport in this region and will make this the subject of a further study.

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References

Angell, J. K., Comparison of stratospheric warming following Agung, El Chichon, and Pinatubo volcanic eruptions, *Geophys. Res. Lett., 20,* 715–718, 1993.

Barath, F., et al., The Upper Atmosphere Research Satellite Microwave Limb Sounder Instrument, *J. Geophys. Res., 98,* 10,751-10,762, 1993.

Brewer, A. M., Evidence for a world circulation provided by the measurements of helium and water vapor distribution in the stratosphere, *Q. J. R. Meteorol. Soc., 75,* 351–363, 1949.

Carr, E. S., et al., Tropical stratospheric water vapor measured by the microwave limb sounder (MLS), *Geophys. Res. Lett., 22,* 691–694, 1995.

Clark, H. L., R. S. Harwood, P. W. Mote, and W. G. Read, Variability of water vapor in the tropical upper troposphere as measured by the Microwave Limb Sounder on UARS, *J. Geophys. Res., 103,* 31,695–31,707, 1998.

Collimore, C. C., M. H. Hitchman, and D. W. Martin, Is there a quasi-biennial oscillation in tropical deep convection?, *Geophys. Res. Lett., 25,* 333–336, 1998.

Coy, L., and R. Swinbank, Characteristics of stratospheric winds and temperatures produced by data assimilation, *J. Geophys. Res., 102,* 25,763–25,781, 1997.

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

Dessler, A. E., A reexamination of the "stratospheric fountain" hypothesis, *Geophys. Res. Lett., 22,* 4165–4168, 1998.

Dessler, A.E, Reply, *Geophys. Res. Lett., 26,* 2739, 1999.

Dickenson, R. E., On the excitation of and propagation of zonal winds in an atmosphere with Newtonian cooling, *J. Atmos. Sci., 25,* 269–279, 1968.

Eliassen, A., Slow thermally or frictionally controlled meridional circulation in a circular vortex, *Astrophys. Norv., 5*(2), 19–60, 1951.

Elson, L. S., W. G. Read, J. W. Waters, P. W. Mote, J. S. Kinnersley, and R. S. Harwood, Upper Tropospheric Water vapor from UARS MLS, *Bull. Am. Meteorol. Soc., 101,* 9001–9015, 1996.

Evans, S. J., R. Toumi, J. E. Harries, M. P. Chipperfield, and J. M. Russell III, Trends in stratospheric humidity and the sensitivity of ozone to these trends, *J. Geophys. Res., 103,* 8715-8725, 1998.

Folkins, I., M. Loewenstien, J. Podolske, S. J. Oltmans, and M. Proffitt, 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.

Gage, K. S and G. C. Reid, Longitudinal variations in tropical tropopause properties in relation to tropical convection and El Niño-Southern Oscillation events, *J. Geophys. Res., 92,* 14,197–14,203, 1987.

Gettleman, A., J. R. Holton, and A. R. Douglass, Simulations of water vapor in the lower stratosphere and upper troposphere, *J. Geophys. Res., 105,* 9003-9023, 2000.

Gettleman, A., M. J. Salby, F. Sassi, The distribution and influence of convection in the tropical tropopause region, *J. Geophys. Res., submitted,* 2001.

Gettleman, A., W. J. Randel, S. Massie, F. Wu, El Niño as a natural experiment for studying the tropical tropopause region, *J. of Clim.,* in press, 2001.

Highwood, E. J., and B. J. Hoskins, The tropical tropopause, *Q. J. R. Meteorol. Soc., 124,* 1579–1604, 1998.

Holton, J. R., P. H. Haynes, M. E. McIntyre, A. R. Douglas, R. B. Rood, and L. Pfister, Stratosphere–troposphere exchange, *Rev. of Geophys., 33*(4), 403–439, 1995.

Holton, J. R., and A. Gettleman, Horizontal transport and the dehydration of the stratosphere, *Geophys. Res. Lett., 28,* 2799-2803, 2001.

Jackson, D. R., S. J. Driscoll, E. J. Highwood, J. E. Harries, and

J. M. Russell III, Troposphere to stratosphere transport at low latitudes as studied using HALOE observations of water vapor 1992–1997, *Q. J. R. Meteorol. Soc.*, *124*, 169–192, 1998.

Labitzke, K., and M. P. McCormick, Stratospheric temperature increases due to Pinatubo aerosols, *Geophys. Res. Lett.*, *19*, 207–210, 1992.

Mergenthaler, J. L., A. E. Roche, J. B. Kumer, and G. A. Ely, Cryogenic Limb Array Etalon Spectrometer observations of tropical cirrus, *J. Geophys. Res.*, *104*, 22,183–22,194, 1999.

Mote, P. W., K. H. Rosenlof, J. S. Holton, R. S. Harwood, and J. W. Waters, Seasonal variations of water vapor in the tropical lower stratosphere, *Geophys. Res. Lett.*, *22*, 1093–1096, 1995.

Mote, P. W., K. H. Rosenlof, M. E. McIntyre, E. S. Carr, J. C. Gille, J. S. Holton, J. S. Kinnersley, H. C. Pumphrey, J. M. Russell III, and J. W. Waters, An atmospheric tape recorder: The imprint of tropical tropopause temperatures on stratospheric water vapor, *J. Geophys. Res.*, *101*, 3989–4006, 1996.

Mote, P. W., H. C. Pumphrey, and T. J. Dunkerton, Sub-seasonal variations in lower stratospheric water vapor, *Geophys. Res. Lett*, *25*, 2445–2448, 1998a.

Mote, P. W., T. J. Dunkerton, M. E. McIntyre, E. A. Ray, P. H. Haynes, and J. M. Russell III, Vertical velocity, vertical diffusion, and dilution by midlatitude air in the tropical lower stratosphere, *J. Geophys. Res.*, *103*, 8651–8666, 1998b.

Mote, P. W., H. L. Clark, T. J. Dunkerton, R. S. Harwood, and H. C. Pumphrey, Intraseasonal variations of water vapor in the tropical upper troposphere and tropopause region, *J. Geophys. Res.*, *105*, 17,457-17,470, 2000.

Newell, R. E., and S. Gould-Stewart, A stratospheric fountain? *J. Atmos. Sci.*, *38*, 2789–2796, 1981.

Pfister, L., K. R. Chan, T. P. Bul, S. Bowden, M. Legg, B. Bary, K. Kelly, M. Proffitt, and W. Starr, Gravity waves generated by a tropical cyclone during the STEP Tropical Field Program: A case study, *J. Geophys. Res.*, *98*, 8611–8638, 1993.

Plumb, R. A., A "tropical pipe" model of stratospheric transport, *J. Geophys. Res.*, *101*, 3957–3972, 1996.

Potter, B.E., and F.R. Holton, The role of monsoon convection in the dehydration of the lower tropical stratosphere, *J. Atmos. Sci.*, *52*, 1034–1050, 1995.

Pumphrey, H. C., Validation of a new prototype water vapor retrieval for UARS MLS, *J. Geophys. Res.*, *104*, 9399-9412, 1999.

Pumphrey, H. C., H. L. Clark, and R. S. Harwood, Lower stratospheric water vapor measured by UARS MLS, *Geophys. Res. Lett.*, *27*, 1691-1694, 2000.

Randel, W. J., F. Wu, and D. J. Gaffen, Interannual variability of the tropical tropopause derived from radiosonde data and NCEP reanalyses, *J. Geophys. Res.*, *105*, 15,509–15,523, 2000.

Reber, C. A., The Upper Atmosphere Research Satellite (UARS), *Geophys. Res. Lett.*, *20*, 1215–1218, 1993.

Reid, G. C., and K. S. Gage, Interannual variations in the height of the tropical tropopause, *J. Geophys. Res.*, *90*, 5629–5635, 1985.

Rind, D., E.-W. Chiou, W. Chu, S. Oltmans, J. Lerner, J. Larsen, M. P. McCormick, and L. McMaster, Overview of the Stratosphere Aerosol and Gas Experiment II water vapor observations: Method, validation, and data characteristics, *J. Geophys. Res.*, *98*, 4835–4856, 1993.

Russell, P. B., L. Pfister, and H. B. Selkirk, The Tropical Experiment of the Stratosphere-Troposphere Exchange Project (STEP): Science objectives, operations and summary findings, *J. Geophys. Res.*, *98*, 8563–8589, 1993.

Schauffler, S. M., and J. S. Daniel, On the effects of stratospheric circulation changes on trace gas trends, *J. Geophys. Res.*, *99*, 25,747–25,754, 1994.

Sherwood, S. C., A "stratospheric drain" over the maritime continent, *Geophys. Res. Lett.*, *27*, 677–680, 2000.

Sherwood, S. C., and A. E. Dessler, On the control of stratospheric humidity, *Geophys. Res. Lett.*, *27*, 2513–2516, 2000.

Sherwood, S. C., and A. E. Dessler, A model for transport across the tropical tropopause, *J. Atmos. Sci.*, *58*, 765–779, 2001.

Swinbank, R., and A. O'Neill, A stratosphere-troposphere data assimilation system, *Mon. Weather Rev.*, *122*, 686–702, 1994a.

Swinbank, R., and A. O'Neill, Quasi-biennial and semi-annual oscillation in equatorial wind fields constructed by data assimilation, *Geophys. Res. Lett.*, *21*, 2099–2102, 1994b.

Trenberth, K. E., The definition of El Niño, *Bull. Am. Meteorol. Soc.*, *78*, 2771–2777, 1997.

Vömel, H., and S. J. Oltmans, Comment on "A reexamination of the 'stratospheric fountain' hypothesis" by A. E. Dessler, *Geophys. Res. Lett.*, *26*, 2737–2738, 1999.

Wang, P.-H., P. Minnis, M. P. McCormick, G. S. Kent, and K. M. Skeens, A 6-year climatology of cloud occurrence frequency from Stratospheric Aerosol and Gas Experiment II observations (1985–1990), *J. Geophys. Res.*, *101*, 29,407–29,429, 1996.

Waters, J. W. Microwave Limb Sounding, in *Atmospheric Remote Sensing by Microwave Radiometry,* edited by M. A. Jansen, pp. 383–496, John Wiley, New York, 1993.

Waugh, D. W., and R. A. Plumb, Contour advection with surgery: A technique for investigating finescale structure in tracer transport, *J. Atmos. Sci.*, *51*, 530–540, 1994.

Zhou, X.-L., M. A. Geller, and M. Zhang, Cooling trend of the tropical cold point tropopause temperatures and its implications, *J. Geophys. Res.*, *106*, 1511–1522, 2001.

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