The water vapour budget of the stratosphere studied using LIMS and SAMS satellite data

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SUMMARY

Monthly zonal mean observations of H₂O and CH₄ made by the limb infrared monitor of the stratosphere (LIMS) and the stratospheric and mesospheric sounder (SAMS) instruments on Nimbus 7 have been used to investigate whether the H₂O mixing ratios in the stratosphere are consistent with a source via the oxidation of CH₄. While both sets of data show considerable seasonally varying structure, total hydrogen (neglecting molecular hydrogen) is relatively featureless with a mean value over the stratosphere of 6.0±0.35 p.p.m.v. (1σ) for the five-month period studied. The uniformity of the total hydrogen fields points to the validity of the CH₄ oxidation hypothesis.

The derived fields of total hydrogen are used to deduce a mean H₂O mixing ratio for air as it enters the stratosphere of 2.7±0.35 p.p.m.v. (1σ) from which a deactivation temperature may be deduced (for example, −87.2°C at 60 mb).

1. INTRODUCTION

Despite the general acceptance of the Brewer hypothesis (1949) that air enters the stratosphere after being freeze-dried at the tropical tropopause, there has in recent years been considerable uncertainty surrounding a number of key areas of the stratospheric water vapour budget. For example, the physical processes that transport water vapour into the stratosphere are uncertain and are the subject of keen debate. Once in the stratosphere the chemical and physical sources and sinks of H₂O are not satisfactorily characterized. In particular, it is unclear whether the increase in H₂O in the upper stratosphere is adequately explained by methane oxidation.

More recently, however, a number of quite revealing observations have been made. The presence of a water vapour minimum above the tropical tropopause at around 19 km (∼60 mb) had been noted by Kley et al. (1979, 1982) using an in situ sampler. The spatial extent and robustness of this feature (the so-called ‘hygropause’) has been demonstrated by satellite observations (Russell et al. 1984a) and it is widely held that understanding the processes that create and maintain the hygropause is central to understanding the cross-tropopause transport of water vapour in tropical latitudes.

Regarding the fate of H₂O once in the stratosphere, measurements by the LIMS (limb infrared monitor of the stratosphere) instrument on Nimbus 7 have established the vertical and latitudinal gradients of H₂O with confidence for the first time. These observations together with simultaneous CH₄ observations from the SAMS (stratospheric and mesospheric sounder) instrument on the same satellite allow much more thorough investigation of the stratospheric water vapour budget than hitherto possible. In this

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paper we use the satellite data to consider this problem, concentrating on the source of \( \text{H}_2\text{O} \) from \( \text{CH}_4 \) oxidation. Some insight into the factors that control the cross-tropopause flux of \( \text{H}_2\text{O} \) can also be gained.

In the next section the salient features of our present understanding of the stratospheric water vapour budget will be reviewed in somewhat more detail. The LIMS and SAMS instruments which provide the measurements of \( \text{H}_2\text{O} \) and \( \text{CH}_4 \) are discussed briefly. We then describe a study which uses the satellite data to calculate the total hydrogen burden of the stratosphere held in methane and water vapour, and hence to study the methane oxidation hypothesis. Finally, the mixing ratios of total hydrogen so derived can also be used to make some inferences concerning the transfer of water vapour from troposphere to stratosphere.

2. THE STRATOSPHERIC WATER VAPOUR BUDGET

(a) Troposphere/stratosphere exchange

In this section we attempt to summarize the water vapour budget of the middle atmosphere, the two main areas of interest being the transport of \( \text{H}_2\text{O} \) from the troposphere to the stratosphere and the \textit{in situ} sources and sinks of stratospheric \( \text{H}_2\text{O} \). The aim is not to provide an exhaustive review but to provide a context for the subsequent discussions.

On the basis of the low humidity found in the low stratosphere, Brewer (1949) proposed an atmospheric circulation with air rising at tropical latitudes, ascending slowly through the cold tropical tropopause where it is freeze-dried, and with generally poleward and descending motion elsewhere. Such a circulation is qualitatively consistent with the observed high latitude ozone distribution (Dobson 1956).

Although it was quickly realized (see e.g. the review by Holton (1984) and references therein) that the Brewer circulation was not a complete description of stratospheric tracer transport and that, particularly in mid-latitudes, large-scale eddies played a major role, the basic concept of desiccation by freeze-drying during injection has remained the sole plausible explanation of the dryness of the stratosphere. However, several problems with the Brewer mechanism emerged. First, measurements by Mastenbrook (1974) and Kley \textit{et al.} (1979) showed lower mixing ratios in the low stratosphere than would be explained by the mean tropical tropopause temperature, or even those temperatures found in the more active convective regions of the intertropical convergence zone. Moreover, the thick, uniform cirrus over tropical latitudes, resulting from a slow mean ascent, was not observed (e.g. Kerley 1960). Second, the measurements of Kley \textit{et al.} showed a minimum in water vapour mixing ratio \( \sim 2-3 \) km above the tropopause. The first problem prompted Newell and Gould-Stewart (1981) to refine the original Brewer hypothesis by proposing that air enters the stratosphere only in restricted regions and at specific times of the year. Defining these as being those areas sufficiently cold to freeze-dry the injected air to \( \sim 3 \) p.p.m.v. at 100 mb, Newell and Gould-Stewart identified the most favoured locations for this to occur as being over the western tropical Pacific, northern Australia, Indonesia and Malaysia between November and March, and over the Bay of Bengal during the monsoon. Newell and Gould-Stewart termed these regions 'stratospheric fountains'. Somewhat similar conclusions were reached by Atticks and Robinson (1983) who identified other tropical locations as also being suitable. Indeed, the climatology of Goldie \textit{et al.} (1958) shows suitable temperatures extending from the Indonesian region as far as East Africa and covering most of the Indian Ocean. However, the fountain hypothesis also has several shortcomings. Firstly, it is not clear that the assumption implicit in the
Newell and Gould-Stewart hypothesis, that no ice or water droplets formed during the ascent will remain in the stratosphere, will be realized. Those ice crystals remaining in the stratosphere would hydrate the stratosphere on evaporation. It should be mentioned that the converse view (that penetrating clouds may actually dehydrate the stratosphere, by acting as a ‘cold finger’) has been put forward by Johnston and Solomon (1979). Secondly, as it stands the fountain hypothesis is not obviously consistent with the observations of Kley et al. (1979) made over Panama in September. It is possible to reconcile the two to some extent by supposing that air over Panama at ~19 km with mixing ratios of about 3 p.p.m.v. originated in the Indonesian region and that, because of rapid zonal flow and slow vertical mixing on a global scale, it has preserved its identity until observed over Panama. However, the Panama observations also showed that air with H₂O mixing ratios in excess of 6 p.p.m.v. (twice that postulated by Newell and Gould-Stewart) had penetrated the tropopause.

A possible dehydration mechanism was proposed by Danielsen (1982). A well-known feature of tropical convective systems is the appearance on short time scales (5–10 hours) of large-scale cirrus anvil clouds (≥200 km) following the overshooting of cumulonimbus turrets. Aerosol measurements by Knollenberg et al. (1982) demonstrated that the overshooting turrets entrain stratospheric air and could ultimately consist of as much as 40% air from the stratosphere. Danielsen pointed out that the effect of this would be to raise the mean potential temperature of the ascending air mass causing anvil formation well above the neutral buoyancy level of an isolated air parcel of surface origin. Given favourable surface conditions and background temperature profiles, anvils could then form in the low stratosphere. Danielsen's suggestion offers an explanation of the low stratosphere temperature profiles characterized by two minima, observed by, for example, Kley et al. (1982) and Johnson and Kriete (1982). Further weight is given to Danielsen's proposed mechanism by the measurements of Kley et al. over Panama who found water vapour mixing ratios at the second temperature minimum equal to the saturation mixing ratio over ice. It should be appreciated, though, that in this region the anvils are found at around 17 km compared with the observed tropical water vapour minimum at ~19 km. Danielsen suggested that the same mechanism applied to other regions and in other seasons could produce higher, colder anvils and lower water vapour mixing ratios would result. An examination of climatological temperatures revealed the most favourable areas and times for this as being precisely those of the 'stratospheric fountain' defined by Newell and Gould-Stewart (1981). However, while Danielsen's hypothesis appears to explain the Panama data, it is yet to be tested in the Indonesian region during the appropriate seasons.

(b) Methane oxidation

The discovery by Meinel (1950) of hydroxyl bands in the airglow and the identification by Migeotte (1948) of CH₄ lines in a solar spectrogram prompted the first research on the photochemistry of H₂O (Bates and Nicolet 1950) and CH₄ (Bates and Witherspoon 1952) in the upper atmosphere. These papers recognized the importance of CH₄ as a source of hydrogen in the upper atmosphere. Later, Bates and Nicolet (1965) noted the role of methane oxidation in the water budget. While the details of the methane oxidation cycle are poorly understood, each stratospheric methane molecule oxidized is thought ultimately to yield approximately two water molecules and hence the net H₂O flux is downward in the stratosphere. Nicolet (1970) recognized that, given representative mixing times in the stratosphere, the oxidation of methane should give rise to increasing H₂O mixing ratios above about 35 km.
The principal chemical sink of H₂O in the stratosphere is reaction with O(1D) with a time constant of many months in the upper stratosphere. While this reaction is extremely important for initiating hydrogen radical chemistry in the stratosphere, it represents an insignificant sink of H₂O. In the mesosphere, photolysis of H₂O by Ly-α radiation and in the Schumann–Runge bands and continuum becomes important and the H₂O mixing ratio is expected to decrease towards the mesopause. Products following the photolysis will include H (ultimately lost to space), OH, HO₂ and, after radical recombination, H₂.

Observations of H₂O in the stratosphere have been reviewed by Harries (1976), Robinson (1980), Ellsaesser et al. (1980), Ellsaesser (1983) and NASA (1979). Two particular problems are evident. Different observations show quite different vertical profiles, with some reported values of greater than 10 p.p.m.v. Secondly, in the low stratosphere, these reviews suggested some latitudinal structure with maximum mixing ratios in low latitudes. These observations are difficult to reconcile with methane oxidation as the only stratospheric source of water vapour. For example, if water can enter the stratosphere at equatorial latitudes and there is a source, but no sink, of water in the stratosphere, then it follows that minimum stratospheric mixing ratios should be found just above the equatorial tropopause. The admittedly scanty evidence on latitudinal structure discussed by Ellsaesser et al. (1980) therefore conflicts with the methane oxidation hypothesis. These problems were discussed in their review.

Consider next the vertical distribution of H₂O. If we assume a mixing ratio for CH₄ of 1.5 p.p.m.v. at the tropopause, then the maximum enhancement of H₂O in the stratosphere by methane oxidation should be 3.0 p.p.m.v. Given initial water vapour mixing ratios of around 3 p.p.m.v., measured mixing ratios approaching 10 p.p.m.v. cannot be reconciled with this theory. However, there is such a large spread within the measurements that the high mixing ratios could possibly be regarded simply as erroneous. Indeed, Ellsaesser (1983) speculates that "instrumental accuracy is the sole cause of the increases with altitude that exceed those attributable to methane oxidation".

To attempt to reconcile the conflicting data, an analysis was described in WMO (1982) using only data from those experiments in which profiles of both H₂O and CH₄ were measured simultaneously. If \([\text{CH}_4]_z = [\text{CH}_4]_{\text{base}}, [\text{H}_2\text{O}]_z = [\text{H}_2\text{O}]_{\text{base}}\), represent a mixing ratio at height z then, it is argued, the quantity
\[
R = -2([\text{CH}_4]_z - [\text{CH}_4]_{\text{base}})/([\text{H}_2\text{O}]_z - [\text{H}_2\text{O}]_{\text{base}})
\]
should equal unity. Implicit in this argument is the assumption that a parcel of air entering the stratosphere maintains its identity until it is sampled; no diffusive mixing takes place. Secondly, it is assumed that all the parcels in the profile entered the stratosphere with the same hydrogen content. The WMO analysis indicates problems with the methane oxidation hypothesis. The ratio, \(R\), was considerably less than unity for all the cases considered, indicating an excess of H₂O over that which could be explained by methane oxidation.

Measurement of water vapour profiles in the stratosphere is difficult and much confusion has been generated by the large spread of the data. It is clear that not all of the published data can be reconciled. A unique possibility to study problems of the water vapour budget has been afforded by the measurements from two infrared sensors on Nimbus 7.

3. **SATELLITE OBSERVATIONS OF H₂O AND CH₄**

The LIIMS and SAMS instruments were launched on the Nimbus 7 satellite in late October 1978. The satellite was positioned in a sun-synchronous near-polar orbit which
provided almost global coverage each day for both instruments. Both instruments viewed the earth's limb (see e.g. Gille and Russell 1984 and Drummond et al. 1980), measuring profiles of infrared radiance in a number of different spectral intervals, from which viewing attitude and profiles of temperature and constituent abundances were deduced. Different satellite viewing geometries meant that the LIMS instrument, which measured H2O mixing ratio, obtained measurements between 84°N and 64°S while the SAMS, which measured CH4 mixing ratios, obtained coverage from 70°N to 50°S. Observations of both H2O and CH4 were made between ~20 mb and 1 mb.

The data used in the subsequent analysis extend from January to May 1979 (when the LIMS cryogen was exhausted). The SAMS continued in operation until June 1983. Monthly, zonal mean mixing ratio fields have been used here.

The LIMS H2O cross-sections and SAMS CH4 cross-sections have been discussed in detail in earlier papers (Russell et al. 1984a, Jones and Pyle 1984, respectively); examples of each are given in Figs. 1 and 2. Detailed error budgets are given in the above references and it is sufficient to say here that the estimated uncertainty on monthly mean cross-sections of both constituents is ~±20% at most heights.

4. THE HYDROGEN CONTENT OF THE STRATOSPHERE

The role of methane oxidation in the hydrogen budget of the stratosphere has been studied previously by taking the ratio \( R = -2\Delta [\text{CH}_4]/\Delta [\text{H}_2\text{O}] \), as discussed above. It was decided that this approach is inadequate for a number of reasons. It implicitly assumes that air enters the stratosphere with a constant hydrogen content, otherwise transport processes would invalidate the procedure. If this were not in fact the case, varying profiles of \( R \) would be interpreted incorrectly in terms of additional sources or sinks. Also, the measurements making up each profile will have instrumental errors associated with them. The quantity \( -2\Delta [\text{CH}_4]/\Delta [\text{H}_2\text{O}] \), computed by differencing quantities of similar size, will inevitably be very sensitive to both bias and random errors. Measurement bias errors also introduce the additional difficulty of selecting a base height from which to begin the calculations. While the problems with atmospheric transport could be largely removed by considering the global average of \( R \), this would not make the analysis any less sensitive to errors.

An alternative and more satisfactory approach is to take the sum of total hydrogen, \( \mathcal{H} = 2[\text{CH}_4] + [\text{H}_2\text{O}] + [\text{H}_2] \). This quantity will be less susceptible to instrumental errors. It has the further advantage that the history of the air parcels need not be known and is not implied. A notional air parcel entering the stratosphere with a given total hydrogen content, partitioned in some way, will preserve that value. As the air parcel encounters wind shear, large-scale mixing, or turbulence, it will mix, or become interleaved between other parcels, possibly with different hydrogen contents. In these instances the satellite observations would give average values for the hydrogen content. This averaging does not affect the validity of our approach. Molecular diffusion would tend to degrade the assumption, but this process is unimportant on the scales we are considering. Such an approach, using balloon data sets, has been followed by Pollock et al. (1980) and Rinsland et al. (1984). To begin with we can reasonably neglect changes in H2 (a constant stratospheric mixing ratio of 0.5 p.p.m.v. is often assumed) and treat that term as an offset. We then consider \( \dot{\mathcal{H}} = 2[\text{CH}_4] + [\text{H}_2\text{O}] \) as our measure of total stratospheric hydrogen.

Before showing fields of \( \dot{\mathcal{H}} \), the individual fields are discussed in more detail. Figures 1 and 2 show cross-sections of CH4 and H2O for January 1979. For CH4, Fig. 1 shows a low altitude, low latitude maximum. For H2O (Fig. 2), minimum mixing ratios are found
Figure 1. Monthly mean cross-section of SAMS methane volume mixing ratio (p.p.m.v.) for January 1979.

Figure 2. Monthly mean cross-section of LIIMS water vapour volume mixing ratio (p.p.m.v.) for January 1979.
in the subtropical lower stratosphere, at the hygropause. Notice that the gradient of H₂O in the low stratosphere is such that the mixing ratios increase polewards (as well as with altitude), qualitatively consistent with the methane oxidation theory.

In Fig. 3 CH₄ and H₂O cross-sections for January and March 1979 are overlayed. Fields of species which are dynamically controlled (and in the stratosphere both water
vapour and methane have photochemical time constants of months or longer) should be expected to show a large degree of similarity and, indeed, remarkable agreement is seen in terms of the general structure of the fields. In fact, the fields are anti-correlated because of the opposite gradients of $H_2O$ and $CH_4$. Notice in January the close coincidence of the 5 p.p.m.v. contour for $H_2O$ and the 0-6 p.p.m.v. contour for $CH_4$. In March excellent correspondence is seen again in the shapes of the contours. At high latitudes the agreement is perhaps not quite so good. This is discussed further below.

Figure 4. $H (=2[CH_4] + [H_2O])$ in p.p.m.v. calculated from LIMS and SAMS data. (a) January 1979; (b) March 1979; (c) May 1979.
It should be remembered that the data sets are quite independent, being derived from different instruments employing different measurement techniques. This is further confirmation that features such as the ‘double peak’ in mixing ratio observed in equatorial latitudes (Jones and Pyle 1984) are real, and not artifacts of the retrieval method. Thus, while the effect of measurement errors, and particularly any biases, must be considered, because of the consistency of the individual measurements and the extensive spatial and temporal coverage, the fields of total hydrogen presented below should carry more weight than previous analyses based on individual balloon ascents.

Figure 4 shows the fields of $\dot{H}$ for January, March and May 1979, calculated from monthly zonal mean LIMS and SAMS data. The calculation was carried out on a grid with resolution of $\pi/19$ ($\sim 9.5^\circ$) in latitude and 0-5 pressure scale heights ($\sim 3.5 \text{ km}$) in the vertical. A uniform field with mixing ratios of about 6 p.p.m.v. is seen. Some structure is apparent although the substantial gradients in the individual data sets (Figs. 1 and 2) are not now present. The total uncertainty in $\dot{H}$ is around 1 p.p.m.v. so that this residual structure is possibly not real. There is also a suggestion of a trend with lowest mixing ratios in January and February (not shown here) but, again, the temporal variation seen is less than the estimated error.

The fact that $\dot{H}$ is apparently a conserved quantity points to CH$_4$ and H$_2$O describing a family in which methane oxidation is the sole stratospheric source of water vapour. Thus the increases in H$_2$O seen in the upper stratosphere by LIMS are consistent with methane oxidation.

Further credence is given to this aspect of photochemical theory if H$_2$ is considered as the additional component of total stratospheric hydrogen, $\mathcal{H}$. The mixing ratio of H$_2$ is about 0.5 p.p.m.v. in the stratosphere and is often assumed to be uniform. However, as discussed previously, H$_2$O is photolysed in the mesosphere with subsequent production of H$_2$. Using their two-dimensional model of the stratosphere and mesosphere, Solomon

Figure 4(c)—continued.
and Garcia (private communication) have shown that air rich in H₂ and poor in H₂O can be carried down into the stratosphere in the polar night region. Figure 5 overlays their model calculation of H₂ for winter with \( \dot{H} \), from SAMS and LIMS, for January. Considering the modelled H₂ fields, some decrease in \( \dot{H} \) at high latitudes might be expected if \( \dot{H} \) is conserved. The minimum in \( \dot{H} \) seen at around 1 mb at 60°N in January (Fig. 4(a)) might be an indication of just this transport from mesosphere to stratosphere with the assumption of constant molecular hydrogen mixing ratios being inadmissible in this region. The May minimum in \( \dot{H} \) around 60°S, 1 mb (Fig. 4(c)) could have a similar cause.

The consistency of the individual data sets and the uniformity of the total hydrogen fields leads us to conclude that the LIMS and SAMS satellite observations of CH₄ and H₂O support the methane oxidation hypothesis. Although this study has used monthly and zonal mean fields of CH₄ and H₂O, similar conclusions were reached in a case study using daily zonal mean values.

It should be mentioned that Remsberg et al. (1984) have also studied the methane oxidation chain using LIMS and SAMS data. They chose to study the ratio \( \Delta[\text{CH}_4]/\Delta[\text{H}_2\text{O}] \) (see above) and concluded that the observations in the upper stratosphere are consistent with methane oxidation as a source of water vapour. Remsberg et al. argue that the loss of one molecule of methane may not lead correspondingly to two water vapour molecules. They point out that photolysis of formaldehyde, an intermediate in the oxidation chain, can lead to production of H₂ (which may be oxidized to form H₂O). However, it seems unlikely that H₂ is a major product of the oxidation chain. Firstly, reaction with OH, leading to H₂O production, should be the main fate of formaldehyde in the upper stratosphere. Furthermore, balloon data (see WMO 1982) do not support an increase of H₂ with height in the stratosphere and indeed suggest, if anything, the contrary. Nevertheless, we feel that the accuracy of the satellite data is not sufficient to

Figure 5. Observed \( \dot{H} \) (p.p.m.v.) for January 1979 (thick lines), and winter H₂ (thin lines) computed by Solomon and Garcia (private communication).
distinguish between complete conversion of methane to water vapour and a minor reaction pathway producing some $\text{H}_2$.

5. TEMPERATURE AND THE WATER VAPOUR MIXING RATIO AT THE HYGROPAUSE

In the previous section it was demonstrated that the satellite observations of $\text{CH}_4$ and $\text{H}_2\text{O}$ are consistent with our understanding of the stratospheric water vapour budget. It is interesting to ask whether further information can be obtained from the satellite data sets about the atmospheric water vapour cycle. In fact, without even specifying the detailed mechanics of the injection process we can, with the data presented above, determine an important parameter, the average water vapour mixing ratio of air as it enters the stratosphere.

In the previous section the sum $2[\text{CH}_4] + [\text{H}_2\text{O}]$ from SAMS and LIMS data was found to be $\sim 6$ p.p.m.v. We know that, at least compared with its stratospheric behaviour, tropospheric methane is well mixed and its mixing ratio well known, with little seasonal variation (see e.g. Ehhalt et al. 1983). Taking the simplest approach, it is possible to

![Figure 6. Distribution of the inferred monthly water vapour mixing ratio (p.p.m.v.) at injection into the stratosphere. Monthly mean values are shown by (--- --). The mean value of the January to May data is also indicated (--- ---).](image)
deduce the water vapour mixing ratio of air during injection from

$$ [\text{H}_2\text{O}]_{\text{injection}} = (2[\text{CH}_4] + [\text{H}_2\text{O}])_{\text{stratosphere}} - (2[\text{CH}_4])_{\text{troposphere}}. $$

Following this approach and taking the tropospheric $\text{CH}_4$ mixing ratio (assumed equal to the injection mixing ratio) to be 1.65 p.p.m.v., we obtain, at the grid points, a mean water vapour mixing ratio for the five-month period of around 2.7 p.p.m.v. for air as it entered the stratosphere. In Fig. 6 are shown the distributions of water vapour injection mixing ratios derived from the SAMS and LIMS observations averaged month by month from January to May.

The small variation evident in the mean mixing ratio from month to month is not statistically significant. Even if it were accepted as real it would be difficult to interpret its significance for troposphere/stratosphere exchange since the air parcels in each monthly average will have a wide range of histories of injection into the stratosphere and transport within it.

The data can nevertheless be used to gain at least one more piece of information. It is possible without a detailed knowledge of the injection mechanism, and given certain assumptions, to deduce the minimum temperature which all air entering the stratosphere must have experienced. If the desiccation level is known, and air is taken to be freeze-dried to its saturation vapour pressure with respect to ice with all water not in the vapour phase falling back into the troposphere, it is possible from a knowledge of the saturation vapour pressure characteristic to derive equivalent temperatures, $T_e$ (the temperature for saturation at the injection pressure), from the SAMS and LIMS data. In Fig. 7 is shown the distribution of this quantity for January to May, derived assuming that desiccation occurs at 100 mb (typical of the tropical tropopause level). It should be noted that the scatter of the equivalent 100 mb $T_e$ of $\pm 0.8^\circ\text{C}$ in this case may not accurately represent the real variability in injection temperature (or water vapour mixing ratio) as

![Figure 7](image_url)

**Figure 7.** Equivalent temperature for injection, assumed to be at the 100 mb level, over the period January to May 1979. The mean value for the period is $-84.2^\circ\text{C}$. 
the spatial averaging inherent in a limb-sounding measurement technique and the monthly and zonal averaging we have used tend to reduce the scatter. On the other hand, measurement errors tend to do the reverse. The scatter in injection temperature will also depend to some small extent on the desiccation level assumed.

However, in view of the tropical water vapour minimum observed at ~19 km by Kley and apparent in the LIMS H\textsubscript{2}O data, it is far from clear that taking 100 mb as the desiccation level is appropriate. In Fig. 8 is shown the variation of the desiccation temperature with desiccation level; it can be seen (curve (a)) that if the drying process occurs at 60 mb (~19 km) the temperature required to give the appropriate mixing ratio is now below ~87°C.

A further consideration which arises if the desiccation level is within the stratosphere rather than at the tropopause is that, either through photochemical destruction or by mixing with air poor in CH\textsubscript{4} (i.e. upper stratospheric air) the methane mixing ratio in the parcel is less than its tropospheric value. In this case the water vapour injection mixing ratio and the equivalent temperature are altered. For example, in Fig. 8, curve (b) shows the relationship between $T_s$ and pressure level given a mean methane mixing ratio of 1.45 p.p.m.v. (approximately that observed by Bush et al. (1978) at 4.5°S at the hygropause level). In this case, $T_s$ at 60 mb becomes ~86.4°C.

6. DISCUSSION

The arguments presented above do not define the detailed physical processes that transfer air into the stratosphere. They do, however, make it clear that the relatively

![Figure 8. Desiccation temperature (°C) against desiccation pressure (mb) calculated from the period-mean water vapour injection mixing ratio. (a) Assuming 1.65 p.p.m.v. of CH\textsubscript{4} at injection; (b) assuming 1.45 p.p.m.v. of CH\textsubscript{4} at injection.](image-url)
humid air (~6 p.p.m.v.) observed over Panama at the altitude of the anvil tops (see Kley et al. 1982) must subsequently pass through a drying region (perhaps the Indonesian region), be mixed with much drier air, or re-enter the troposphere, for example, as indicated by the model experiment of Kida (1983).

An interesting question is whether the equivalent temperatures referred to above can themselves be used to identify the geographical locations where injection might occur. At first sight the 100 mb equivalent $T_e$ values point very closely to the Indonesian region as being responsible, particularly in the early months of the year, as suggested by Newell and Gould-Stewart (1981). However, if the Danielsen hypothesis is accepted, such an argument may not be entirely satisfactory even if climatological temperatures representative of the hygropause region rather than 100 mb were used. This is because the essence of Danielsen's hypothesis is that the temperature profile is perturbed for a relatively short period (~10 hours). This, and the small spatial extent of the drying region, will almost certainly mean that the climatological temperatures obtained from large numbers of sonde ascents will not be typical of anvil conditions. Rather, the low 100 mb temperatures seen for example in the Indonesian region are merely indications of the presence of the deep convection that is essential for anvil formation at the required levels.

For similar reasons a satellite-derived temperature climatology for the hygropause may not be particularly useful. Satellite measurements of the disturbed (anvil) events are anyway difficult to interpret since the presence of cloud perturbs the emitted radiances. The LIMS procedure was to ignore such contaminated measurements, a process which itself could lead to biases in the derived temperatures and mixing ratios. In the event, below 70 mb the LIMS temperatures show clear positive biases (Gille and Russell 1984) implying the likelihood of compensating negative bias errors in the LIMS-derived water vapour fields at and below this level. Other, in situ, measurements of water vapour in the low stratosphere show significant and as yet unexplained differences which are at least as large as the anticipated LIMS bias errors. To bypass this uncertainty we have therefore chosen not to derive saturation temperatures by direct measurement (either of temperature or water vapour mixing ratio), but by using mixing ratio data from above 20 mb where no problems of this kind occurred.

Given the existence of the hygropause it is relevant to ask how numerical models, without producing a hygropause or comparable feature and without a detailed description of the $\text{H}_2\text{O}$ transport mechanism, have managed to produce realistic stratospheric water vapour distributions. Clearly, the answer is that at some pressure the modelled temperature must (at all longitudes) have coincided with some point on the desiccation curve (or its extrapolation) shown in Fig. 8. The range of suitable pressures and temperatures is wide and it is no surprise that the coincidence occurs in some models.

7. CONCLUSION

LIMS and SAMS data on stratospheric water vapour and methane have been used to investigate some features of the water vapour budget of the middle atmosphere. Comparison of the individual fields points to a very high degree of consistency between the data sets and, incidentally, gives additional confidence to other studies using these data. By considering $\dot{H} = 2(\text{CH}_4) + [\text{H}_2\text{O}]$ it was demonstrated that this appears to be a conserved quantity and that methane oxidation is the reason for the increase of water vapour seen in the upper stratosphere. We would agree with Ellsaesser (1983) that water vapour profiles showing increases with altitude which exceed those attributable to methane oxidation are probably due to the problems of inadequate instrumental accuracy.
There is a suggestion in the data that during certain months H₂O-poor air is brought down into the stratosphere in the polar night region. This would be consistent with the transport of NOₓ discussed by Russell et al. (1984b).

The data can be used to infer an average water vapour mixing ratio on injection into the stratosphere of ~2.6 to 2.8 p.p.m.v. This compares with ~3.4 p.p.m.v. observed by Kley (paper presented at NASA meeting, Starnberger See, 1984) and values of less than 2 p.p.m.v. sometimes observed by LIMS. As mentioned above, the LIMS data at the hygropause level are thought to be biased low. While both values are consistent to within their experimental errors, some of the difference between our results and those of Kley could be due to our assumption of 1.65 p.p.m.v. for the CH₄ mixing ratio at the hygropause. A smaller value, implying some mixing or photochemical destruction prior to desiccation, would reduce the discrepancy.

An equivalent temperature required to freeze-dry air to the derived injection mixing ratios of water vapour was determined to be approximately −87°C at 60 mb. However, without a detailed knowledge of the injection mechanism, it is not possible to relate the equivalent temperature to any observed climatological variables (e.g. 100 mb temperatures). For this reason, the identification by Newell and Gould-Stewart (1981) of the fountain regions on the basis of climatological 100 mb, or tropopause, temperatures is questioned. It does appear (Danielsen 1982) that the Indonesian region and the western tropical Pacific are favourable for the development of deep convection essential for the formation of cirrus anvils at the hygropause level, although it is pointed out that suitable conditions may well extend over a wider area.

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