Radiometric Inference of Stratospheric Water Vapor

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ABSTRACT

By varying the amount of water vapor as input to the radiative power transfer equation, assuming a constant carbon dioxide and varying ozone distribution, it is possible to infer stratospheric water vapor from broadband observations of downward irradiance. The procedure is iterative in that downward observed and calculated irradiances, at several levels for each of several radiometric soundings, are brought within the limits of a convergence criterion. This is accomplished by successively reducing an initial overestimate of the stratospheric mixing ratio, defined by a power law, until the sum of the squared differences of observed and calculated irradiances is minimized. The sum includes all levels of the sounding.

Results for a continental area during winter months indicate that the stratospheric water vapor content from 50 mb upward to 10 mb decreases from approximately 20 to 3 parts per million. For tropical Guam and Canton Island the corresponding magnitudes are larger, decreasing from 21 to 4 ppm. The standard deviation of the mean for all pressure levels is approximately 1.0 ppm. Adding deviation to the values inferred should give an upper bound to the water vapor content. The average mixing ratio for the continental stations between 25 and 10 mb is 5.7 ppm with a standard deviation of the mean of 0.8 ppm. Since the infrared radiative emission and attenuation of aerosols is inseparable from emission and attenuation of the atmospheric gases when measured with a broad response radiometer, these mixing ratio results would be reduced by the presence of aerosols. In view of apparent aerosol contamination we have made no inferences below 50 mb (21 km). The results may be said to be an upper bound to the actual quantity of water vapor, favoring an increasingly dry stratospheric profile.

List of Symbols

Δw Change in mixing ratio (parts per million)
ρ Pressure
T Temperature
0 Subscript denoting surface
θ Angle (radians)
ν Wave number (reciprocal centimeters)
τ Transmissivity
I_b Average black body radiance for layer
w Subscript denoting water vapor
c Subscript denoting carbon dioxide
o Subscript denoting ozone
I_b Black body radience
F Downward irradiance (cal cm⁻² min⁻¹)
F'0 Observed downward irradiance (cal cm⁻² min⁻¹)
F'1e Calculated downward irradiance
W Water vapor mixing ratio (ppm)
ρ₀ Pressure at reference level, 50 mb.
λ Exponent describing lapse rate of w(ρ)
W(ρ₀) Mixing ratio at reference level ρ₀
φ Mean water vapor mixing ratio
N Number of pressure levels at which F'1e is calculated
du Increment of optical mass
g Acceleration due to gravity

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1. Introduction

In a recent study, Mastenbrook (1966) presented the results of frost point observations of moisture in the lower stratosphere. His measurements show that the lower stratospheric and upper tropospheric mixing ratio ranges from 2.2–3.2 ppm. A prior publication by Mastenbrook (1965) showed a mixing ratio range of from 1.0–4.0 ppm. The latter work generally agrees with independent spectral measurements and analyses of Houghton et al. (1961) and Calfee and Gates (1966). Murcray et al. (1966), however, indicate a decrease of mixing ratio from approximately 30 through 15 mb with a subsequent increase presumably due to balloon and train contamination. This could also explain Mastenbrook's (1966) results at terminal altitude.

Our purpose is to describe a relatively low-cost radiometric technique for estimating a profile of stratospheric water vapor and to present results of the employment of this technique upwards from 50 mb. Kostyanoy (1965) inferred total water vapor distribution above 100 mb using a similar method on radiometric irradiance data observed over Russia.

In this study over 200 direct observations of downward irradiance are data for a solution of the radiative transfer equation. Downward irradiance calculations are adjusted to reproduce the measured downward irradiance by varying the complete stratospheric water vapor profile. This is accomplished by a least squares
method of minimizing the difference between observed and calculated irradiances. When a predetermined irradiance convergence criterion is reached, a unique solution exists.

The basic assumptions for the calculations are:

1) The background for the downward irradiance is zero at 0.1 mb.
2) The carbon dioxide is distributed uniformly at 0.031 percent by volume.
3) The total optical mass of ozone above 50 mb is 0.212 cm (Craig, 1965). The distribution of ozone is given in Table 1.
4) The presence of optically active aerosols will reduce the inferred water vapor mixing ratios. The average mixing ratio derived may be said to be an average upper bound. Pilipowskyj\(^3\) has shown that radiatively active aerosols may be present as high as 30 mb.
5) The temperature profile from 1.0 to 20.0 mb is taken from Murgatroyd (1957) to fill in high-level temperatures in the downward irradiance calculation. Table 2 gives this profile. It should be noted that a change of ±5.0°C in the 1.0-mb temperature results in a calculated mixing ratio change of only ±0.2 ppm.

The observational requirement is the accumulation of irradiance observations from as high as possible downward through the stratosphere. For this the IQSY radiometer-sonde network provided the necessary data. The accuracy of these measurements is such that the random error in the evaluation of the mixing ratio is 2.2 ppm.

A computer solution for the convergence of calculated to observed irradiance is, of course, a necessity in view of the large number of calculations required. The procedure is one of least squares, beginning with an overestimate of the input water vapor profile for each

\(^3\) Personal communication, S. J. Pilipowskyj, Dept. Meteor., Univ. of Wisc.

### Table 1. Vertical distribution of ozone from 0.1 to 50.0 mb after Craig (1965).

<table>
<thead>
<tr>
<th>Pressure (mb)</th>
<th>Optical depth (mass) increment (gm cm(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>0</td>
</tr>
<tr>
<td>0.3</td>
<td>0</td>
</tr>
<tr>
<td>0.6</td>
<td>0</td>
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<tr>
<td>3.0</td>
<td>0.001</td>
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<tr>
<td>5.0</td>
<td>0.005</td>
</tr>
<tr>
<td>7.0</td>
<td>0.007</td>
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<tr>
<td>10.0</td>
<td>0.010</td>
</tr>
<tr>
<td>15.0</td>
<td>0.020</td>
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<td>20.0</td>
<td>0.030</td>
</tr>
<tr>
<td>25.0</td>
<td>0.030</td>
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<tr>
<td>30.0</td>
<td>0.030</td>
</tr>
<tr>
<td>35.0</td>
<td>0.030</td>
</tr>
<tr>
<td>40.0</td>
<td>0.020</td>
</tr>
<tr>
<td>45.0</td>
<td>0.015</td>
</tr>
<tr>
<td>50.0</td>
<td>0.012</td>
</tr>
</tbody>
</table>

### Table 2. Temperatures between 0.1 and 20.0 mb after Murgatroyd (1957).

<table>
<thead>
<tr>
<th>Pressure level (mb)</th>
<th>Summer, 40N (°C)</th>
<th>Winter, 40N (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>-46.0</td>
<td>-30.0</td>
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<tr>
<td>0.4</td>
<td>-8.0</td>
<td>-5.0</td>
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<tr>
<td>1.0</td>
<td>+12.0</td>
<td>-13.0</td>
</tr>
<tr>
<td>2.0</td>
<td>+4.0</td>
<td>-24.0</td>
</tr>
<tr>
<td>3.0</td>
<td>-4.0</td>
<td>-33.0</td>
</tr>
<tr>
<td>4.0</td>
<td>-12.0</td>
<td>-39.0</td>
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<td>6.0</td>
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<td>8.0</td>
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<td>10.0</td>
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<td>-40.0</td>
<td>-57.0</td>
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<td>16.0</td>
<td>-44.0</td>
<td>-59.0</td>
</tr>
<tr>
<td>20.0*</td>
<td>-47.0</td>
<td>-59.0</td>
</tr>
</tbody>
</table>

*Temperatures corresponding to pressures higher than 20.0 mb are observed for the actual ascent.

atmospheric layer and adjusting the profile downward until the sum of the squared differences of observed and calculated irradiances is reduced to the convergence criterion described in Section 3.

### 2. The observations of atmospheric irradiance

To realize a random error of 2.2 ppm in the radiometrically inferred mixing ratio determined by the rate of the irradiance convergence criterion, it is necessary that the standard deviation of the radiometer-sonde not exceed 0.0012 cal cm\(^{-2}\) min\(^{-1}\). Analyses of radiometer-sonde accuracy have appeared in recent meteorological literature. Kuhn and Johnson (1966) show that the standard deviation of a filtered radiometer-sonde observation due to random error is 0.0012 cal cm\(^{-2}\) min\(^{-1}\) for a 0.2C radiometer-sonde sensor temperature error.

Table 3 illustrates a determination of the random error of a derived mixing ratio for two cases with constant mixing ratio profiles. The table is a computer solution of the radiative power transfer equation. The mixing ratio profiles employed are 1.0 and 4.0 ppm. The average irradiance difference for these two calculations is 0.0016 cal cm\(^{-2}\) min\(^{-1}\). Assuming an instrumental standard deviation of 0.0012 cal cm\(^{-2}\) min\(^{-1}\), the random error in the inferred mixing ratio is 2.2 ppm.

The pressure scaling term for water vapor in both instances is \((p/p_0)^{0.2}\) after Moller and Raschke (1963). Pressure scaling for carbon dioxide is in the form \((p/p_0)^{0.8}\). Elsasser's (1960) water vapor and carbon dioxide flux emissivities were employed. These tables are based on the work of Howard et al. (1955). Temperature scaling in two instances followed the form \((T_0/T)^{0.5}\).

Since the ozone content of the atmosphere above 50 mb can possibly vary by a factor of two, it is necessary to show that such a variation will not obscure the inferred water vapor amounts. Table 4 is an analysis of the contributions of the various gases to the total irradiances at various levels. Two calculations were
made. The water and carbon dioxide profiles were the same in both calculations. The ozone profile for the first calculation was halved at each level. The difference in the radiative contribution of the ozone from the first to the second calculation is given in the column, "\( F_{\mu}^{\bullet \bullet} - F_{\mu}^{\bullet \bullet \bullet} \), ozone." For comparison, the last column of Table 3 lists the difference in the radiative contribution of water vapor in two calculations having constant mixing ratios of 4.0 and 1.0 ppm, respectively. From the last two columns of Table 4, it is apparent that the effects of halving the ozone content are approximately a factor of two less than those radiative effects of a water vapor reduction of 3.0 ppm.

Approximately eighty IFSY radiometric ascents at Guam, Mariana Islands, Canton Island, Green Bay, Wisc., and Washington, D. C., were used in this analysis. All of these ascents exceeded terminal elevations of 15 mb and averaged 6 ± 2 mb at highest altitude. The observations were taken between October 1964 and March 1965.

3. Calculated radiant power and inferred water vapor

Since the irradiance observed by a flat plate radiometer ascending in the atmosphere is related to the atmospheric temperature, pressure, water vapor, carbon dioxide, ozone, and aerosol profiles, the radiative power transfer work (Wark and Fleming, 1966; Elsasser, 1960) uniquely expresses this relationship only in the absence of aerosols. For a plane parallel atmosphere, containing no scatterers, and in local thermodynamic equilibrium, the radiative transfer equation, expressed
in terms of radiance, is given by

\[
I(\theta) = -\int_{\tau_1}^{\tau_2} \int_{1}^{\tau_2} I_b(v, T(\phi))d\tau(v, \phi)d\nu

- \int_{\tau_1}^{\tau_2} \int_{1}^{\tau_2} I_b(v, T(\phi))d\tau(v, \phi)d\nu

- \int_{\nu_1}^{\nu_2} \int_{1}^{\nu_2} I_b(v, T(\phi))d\tau(v, \phi)d\nu

\quad + I_b(v, p_0)\tau(\nu, p_0)\tau(\nu, p_0)\theta. \tag{1}
\]

Irradiance, measured by the radiometersonde, is related directly to radiance, Eq. (1), by the expression

\[
F = 2\pi \int_{0}^{\theta_1} I_b(\theta) \cos \theta \sin \theta d\theta, \tag{2}
\]

where the limit \(\theta_1\) is \(\pi/2\) radians.

Thus, Eq. (2) represents a three-fold integration

1) over all angles \(\theta\),
2) over all frequencies of radiation \(\nu\), and
3) over all transmissivities of the absorbing gas \(\tau\).

Elasasser (1960) has shown that replacing the monochromatic radiance \(I_b(\nu)\) or spectral radiance \(I_b\) by the monochromatic irradiance \(F_b(\nu)\) or by the spectral irradiance \(F_b\) does not change the basic transfer formalism of Eq. (1). Combining Eqs. (1) and (2) and considering only hemispheric downward irradiance, integrated over all wavelengths [fourth term on right hand side in Eq. (1) is zero], the downward irradiance is given by

\[
F_{\downarrow e} = -\int_{1}^{\tau_2} F_b(T(\phi))d\tau(T(\phi))

- \int_{1}^{\tau_2} F_b(T(\phi))d\tau(T(\phi))

- \int_{1}^{\nu_2} F_b(T(\phi))d\tau(T(\phi))d\nu, \tag{3}
\]

where \(F_b\) is the black body irradiance equal to \(\pi I_b\) and \(\tau_T\) is the irradiance transmissivity.

Knowing the irradiance, and temperature profile from 0.1 mb down to approximately 50 mb, we can solve the transfer equation for all levels, assuming a water vapor profile \(w(P)\). The data for the computer program are pressure, mean temperature for each layer, ozone concentration, carbon dioxide concentration, and the assumed water vapor mixing ratio \(w(\phi)\), for each level.

After observing that level-by-level computations of water vapor agreed well with a power function representation of the vapor profile, further water vapor deductions were made assuming a power function represenation given by

\[
w(\phi) = w(p_0)(\frac{\phi}{p_0})^\lambda. \tag{4}
\]

This function was suggested by Smith (1966). It should be noted that the power function approximation can describe an increase in mixing ratio with decreasing pressure when lambda is negative. This possibility was considered; however, it was not possible to make the observed and calculated irradiance values approach agreement using reasonable values for \(w(p_0)\) when lambda was negative. Typically, \(\lambda\) ranged between 0.0 and 1.0, and \(w(p_0)\) between 40.0 and 5.0 ppm. In addition, the calculation of the downward irradiance requires that the water vapor contribution be zero at that pressure where carbon dioxide emission accounts for the entire downward irradiance. The changing of an entire water vapor profile by the power function approximation of Eq. (4) imposes a stabilizing constraint on the solution of Eq. (1) that is not possible in a level-by-level iteration. Moreover, the solution of (1) for water vapor is more efficient when (4) is employed. Initially, overestimation of the profile of \(w\) (gm kg\(^{-1}\)) is made by employing the maximum \(w(p_0)\) and minimum \(\lambda\). The profile is then successively reduced until the quantity \(\sum(F_{\downarrow e} - F_{\downarrow e}^0)^2\) reaches the predetermined convergence criterion, \(N(0.0012)^2\).

The pressure and temperature scaled precipitable centimeters of water vapor and mixing ratio are output for each computed level when convergence occurs. The mean mixing ratio between two successive pressure levels is given by the expression

\[
\bar{w} = d\bar{w} \left(\frac{g}{dP}\right)(\frac{P_0}{\phi})^{0.28} \left(\frac{T}{T_0}\right)^{0.5}. \tag{5}
\]

4. Results

Figs. 1 through 5 are typical profiles of observed irradiance and inferred water vapor mixing ratios in parts per million. The errors in pressure and irradiance are indicated by bars on the plotted points for observed downward irradiances at Green Bay in Fig. 1. This figure also illustrates the least square fitted curve of irradiance for the plotted data points. Such a curve was fitted to each of the 210 ascents employed in the study. However, convergence of calculated-to-actual observed data was accomplished at the 7-, 10-, 15-, 20-, 25-, 30-, 35-, and 50-mb levels. An attempt has been made to clearly compress as much data as possible into each figure and yet retain a measure of simplicity.

A feature of some of the observed irradiance values is the tendency for the magnitude of the highest one or two irradiance observations, immediately prior to balloon burst, to fall below that which carbon dioxide alone contributes. Pressure scaling for the carbon dioxide optical mass is critical at such levels. There is
also the possibility of a systematic error in the instrument at such high altitudes due to a loss of convective temperature coupling of the environment to the polyethylene convection shields. This could cause a small conduction error in the solution of the radiometer equation. However, at pressures greater than 10 mb the convergence criterion of observed and calculated downward radiant power is easily achieved.

The water vapor results illustrated in these figures demonstrate the possibilities of the use of irradiance observations to infer stratospheric water vapor. The examples were not chosen to typify the average profiles listed in Table 5. It should be recalled that we are using the water vapor and carbon dioxide transmissivities of Howard et al. (1955) converted to slab emissivities for
the transfer computations and that the maximum resolution of the radiometrically inferred mixing ratios, due to random errors in the irradiance observations, is 2.2 ppm.

Figs. 1 and 2 are continental profiles of water vapor displaying a decrease in water vapor with height from 35 through 5 mb (pressure error is ±2.0 mb). In Fig. 2, the mixing ratio is constant from 35 through 20 mb. Above this level the decrease in mixing ratio with height, common to the majority of our ascents coupled with the steady decline in the optical mass of carbon dioxide, is reflected in the decrease in the observed and calculated irradiance. This decrease occurs in the presence of a slight increase in temperature with height. The rapidly decreasing concentration of radiant power absorbers and emitters and the zero background radiation counteract the temperature increase. Recent general concern over contamination during ascent of frost point moisture measuring equipment indicates a possible modification of the concept of rapidly increasing mixing ratio above 50 mb.

Figs. 3 and 4 illustrate "wetter" continental mixing ratio profiles that were observed in 25 per cent of the observations. These results show a decrease in mixing ratio with height.

Fig. 5, for Canton Island, demonstrates a typical stratospheric inferred water vapor profile for the tropics above 35 mb. There is no tendency toward a constant mixing ratio. The average sounding in the tropics for this study was considerably more moist than the continental mean sounding.

The first five columns of Table 5 summarize the mean mixing ratios in parts per million, and the standard deviation of the mean for the various station groups and pressure levels. Groupings were arranged to achieve a statistically significant quantity of data and were carried down to 50 mb. The random error of 2.2 ppm in the inferred mixing ratio does not limit estimation of mixing ratio to values larger than 2.2 ppm. It is rather a 68 per cent confidence limit.

Several points are evident from this summary. Primarily, the radiometrically inferred water vapor profile decreases with height. Except for the 10-mb level, the standard deviation of the mean remains approximately constant with height. There is a significant difference in the continental and tropical means for the levels included. Possibly, tropical thermal injection of moisture into the stratosphere results in the mean oceanic soundings displaying more water vapor than the mean continental soundings.

The remaining columns of Table 5 list, for comparison, the measured water vapor mixing ratios for corresponding pressure levels by Mastenbrook (1966), Murcray et al. (1966), and Zander (1966), and the calculations of Caffee and Gates (1966). From 25 through 15 mb, Murcray's data agree quite closely with the continental data of this study. Caffee and Gates show less moisture from 25 through 15 mb than the present study and that of Mastenbrook (1966).

The optical depths for continental stations from 50 through 15 mb reported by this study, by Mastenbrook and by Gates, are approximately 1.2, 1.7, and 1.0, respectively. Mastenbrook's (1966) mixing ratio data at pressures equal to and lower than 20 mb may indicate contamination by moisture from the balloon train. Murcray's results give approximately the same optical depth for the 50-10 mb layer as this study while Zander's results indicate a total optical depth, for the layer discussed, of 4.1 μ.

Balloon-borne radiometer measurements made by Williamson (1964) in the spectral region 5.5-7.0 μ above the tropopause show an essentially constant mix-

### Table 5. Statistical summary and comparison of inferred water vapor (ppm).

<table>
<thead>
<tr>
<th>Station group</th>
<th>Pressure (mb)</th>
<th>Number of observations</th>
<th>Average mixing ratio</th>
<th>Standard deviation of mean</th>
<th>Mastenbrook mixing ratio</th>
<th>Murcray et al. mixing ratio</th>
<th>Caffee and Gates mixing ratio</th>
<th>Zander mixing ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>Washington</td>
<td>10</td>
<td>37</td>
<td>2.7</td>
<td>0.3</td>
<td>12.0</td>
<td>2.8</td>
<td>10.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>15</td>
<td>37</td>
<td>4.9</td>
<td>0.7</td>
<td>3.2</td>
<td>4.9</td>
<td>2.8</td>
<td>10.0</td>
</tr>
<tr>
<td>Green Bay</td>
<td>20</td>
<td>37</td>
<td>6.8</td>
<td>1.0</td>
<td>2.7</td>
<td>6.8</td>
<td>2.8</td>
<td>10.0</td>
</tr>
<tr>
<td></td>
<td>25</td>
<td>35*</td>
<td>8.3</td>
<td>1.2</td>
<td>2.6</td>
<td>8.4</td>
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<td>10.0</td>
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<tr>
<td></td>
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<td>35</td>
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<td>2.3</td>
<td>3.8</td>
<td>2.8</td>
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</tr>
<tr>
<td>Guam, Canton Is.</td>
<td>10</td>
<td>33</td>
<td>4.1</td>
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<td>2.2</td>
<td></td>
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<tr>
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<td>15</td>
<td>33</td>
<td>12.8</td>
<td>1.3</td>
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<td>1.1</td>
<td>2.2</td>
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<tr>
<td></td>
<td>50</td>
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<td>21.2</td>
<td>1.5</td>
<td>2.2</td>
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<td>Continental average</td>
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<td>0.5</td>
<td></td>
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</tr>
<tr>
<td>Tropical average</td>
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<td>131</td>
<td>14.2</td>
<td>0.9</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

* Data missing for some observations.
ing ratio of approximately 3.0 ppm. This corresponds to a total optical depth in the layer 50 through 10 mb of approximately 1.2 μ. Kostyanoy’s (1965) radiometrically inferred water vapor optical depth above 100 mb for 16 radiation sonde ascents over western USSR averaged 35.0 μ. Assuming a constant mixing ratio with height, his results would indicate 14.0 μ in the 50–10 mb layer.

The radiometrically inferred water vapor profile of this research from 50 through 100 mb gives values up to 40 ppm and may indicate broad band aerosol radiative “contamination” at pressures greater than 35–50 mb. These values average higher than most direct measurements for the 50 through 100 mb layer and are not reported on in detail. The increase in deduced water vapor below the 35-mb level suggests that a radiatively effective upper limit for atmospheric aerosols may exist for these radiation soundings at approximately 35 mb.

Fig. 6, after the technique of Gutnick (1961), compares recent measurements and calculations of stratospheric water vapor profiles between 50 and 10 mb by a number of researchers. Mastenbrook’s (1965) data, showing constant mixing ratio with height, curve M1, has been included.

5. Conclusions

Water vapor is the only variable in the radiative power transfer calculation when the distribution of the other gases is known and the temperature profile is known. By varying the amount of water vapor input to the radiative transfer calculation we can successively adjust the water vapor profile to obtain convergence of water vapor calculations with direct observations to a preset limit. This convergence criterion is determined by the random error of the radiometer. Knowing this error, we then infer water vapor quantities from direct observations of irradiance.

Our results, for the winter months, for tropical and mid-latitude continental areas in 1964 and 1965, indicate that the stratospheric water vapor content above 30 mb for continental regions averages 5.6 ppm with a standard deviation of the mean of 0.5 ppm. For the tropics these quantities are 14.2 and 0.9 ppm, respectively. Radiative aerosol “contamination” is a problem below 35 mb as this radiative effect is inseparable from that of water vapor. The radiometrically inferred water vapor mixing ratios do not display a tendency toward a constant mixing ratio with height but rather show a decrease in mixing ratio with height above 50 mb.

REFERENCES


