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On the Structure of the Venus Atmosphere[1]

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We compare the results of the Mariner 5 and Venera 4 spacecraft measurements with one another and with ground-based observations to learn about the structure and composition of the Venus atmosphere. Several independent arguments imply that the carbon dioxide mixing ratio lies between 50 and 85%. Thus some other gas, perhaps nitrogen, must be present in significant amounts. The atmosphere exhibits an adiabatic temperature gradient at temperatures in excess of 400°K and has a slightly subadiabatic value between the 400°K level and a level 15°K warmer than stratospheric temperatures.

The final measurements of Venera 4 appear to have been made not close to the surface, but rather some 25 km above. According to this view, the altitude measurement by the Venera 4 radar altimeter was made when the spacecraft was almost exactly twice the reported 26 km above the surface, and so raises questions as to possible ambiguities in the radar system.

A simple reflecting layer model of line formation yields pressures consistent with polarimetric and space vehicle results; the dependence of the equivalent width on phase angle is not necessarily incompatible with a modified version of this model. Estimates of pressures near the cloudtops from polarization data indicate that there is an appreciable quantity of aerosols in the stratosphere. Finally, the weak water vapor lines observed in the near-infrared are not incompatible with the presence of ice clouds.

Introduction

Due to the remarkable successes of the Venera 4 and Mariner 5 space vehicles, as well as new ground-based results, there is now a wealth of information available relevant to the structure of the Venus atmosphere. The aim of the present paper is to explore the joint implications and mutual consistency of these data. We will investigate the temperature and pressure structure of the atmosphere, the values of these parameters at various boundaries such as the surface, and the carbon dioxide mixing ratio implied by the data. We will also examine the implications of the observed structure of the atmosphere for the problem of absorption line formation, and in particular consider the compatibility of water clouds with the various searches for Venus water vapor lines. We begin by estimating the cloudtop pressure from polarization measurements and the cloudtop temperature from infrared radiometry. Next, the line formation problem is

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considered, and finally we discuss the structure of the lower Cytherean atmosphere.

Pressures Near the Cloudtops

We wish to estimate pressures near the cloudtops from polarization measurements and to compare these values with the tropopause pressures as determined by the Mariner 5 observations. While the atmosphere above the clouds contributes little to the outgoing radiation at almost all detected wavelengths, its contribution is much more highly polarized near phase angle $\Phi = 90°$ than the contribution from light multiply scattered from the cloud layer. Accordingly, it may be possible to determine the amount of atmosphere above the clouds from polarization data. The highly polarized contribution from the atmosphere will consist of photons scattered by the gas out of the atmosphere before they encounter cloud particles, consequent multiple scattering and depolarization. Accordingly, the polarimetric cloud pressures obtained below refer not to the uppermost portions of the clouds, but rather to a cloud optical depth near unity for isotropically scattering cloud particles. A somewhat larger optical depth applies for cloud particles which scatter preferentially into the forward hemisphere.

Relevant formulas for calculating these upper cloud pressures from polarization measurements are presented in Appendix A. One approach for separating out the atmospheric contribution to the observed polarization is to compute the difference in polarization at two wavelengths for $\Phi = 90°$, and to assume that the contribution of the clouds to the polarization is independent of wavelength over the wavelength interval chosen. Under these circumstances Eq. (A3) can be used to find B_a/B_T, the ratio of the atmospheric brightness to the total brightness, for the shorter wavelength. Alternatively, if Rayleigh scattering sufficiently dominates the polarization curve near $\Phi = 90°$, as is clearly indicated by a maximum polarization near this phase angle in the ultraviolet data, a second method may be used to obtain the relative brightness: The difference between the polarization at 90° and the average of the polarization at $\Phi = 90° - \theta$ and $90° + \theta$, where $\theta \ll 90°$, can be attributed entirely to the atmosphere. In this case Eq. (A6) applies. The cloud contribution to the polarization (and the ratio of brightness) is assumed to vary linearly with phase angle between $90° - \theta$ and $90° + \theta$, an assumption amounting essentially to a Taylor series expansion. The polarization data at 9900 Å—a wavelength largely free of Rayleigh scattering—shows, to first approximation, just such a variation for the range of phase angles employed. Equations (A3) and (A6) assume, in addition, that the Rayleigh scattering optical depths are much less than unity, a good assumption in view of the small absolute value of the polarization measured for Venus. These assumptions will be tested by intercomparing the methods and the results at various wavelengths.

Once the ratio of atmospheric to total brightness at $\Phi = 90°$ is obtained, the Rayleigh scattering optical depth may be found from Eq. (A11). This in turn is related to the upper cloud pressures by Eq. (A12). According to Eqs. (A3), (A5), (A6), (A11), and (A12), the upper cloud pressure can be obtained from the polarization measurements, the acceleration of gravity, the geometric albedo, and the ratio of observed fluxes at phase angles 90° and 0°. There is also a weak dependence on the composition of the atmosphere. We employ the polarization measurements of Gehrels and Samuelson (1961), and of Coffeen and Gehrels (1969), the photometric observations of Knuckles, Sinton, and Sinton (1960), and the values for the Cabannes molecular asymmetry depolarization parameter f given by van de Hulst (1952). We assume that the albedo of Venus is approximately constant between 5500 Å and 10000 Å, in agreement with the recent results of Irvine *et al.* (1968), and further assume the atmosphere to be 100% CO_2, a point to which we return below.

Table I presents our resulting estimate of the upper cloud pressures, derived from Eqs. (A3), (A11), and (A12), and based on the $\Phi = 90°$ difference between polarization

TABLE I

UPPER CLOUD PRESSURE DERIVED FROM POLARIZATION MEASUREMENTS AT TWO WAVELENGTHS AND 90° PHASE ANGLE

λ_1 (Å)	λ_2 (Å)	Pressure (mb)
5600	9900	130
5600	6830	127
4200	9900	48
3590	9900	34
3250	9900	29

values at two wavelengths. The pair of wavelengths used is indicated for each pressure estimate. In this method we assume that the polarization attributable to the clouds is the same at the two wavelengths. A check on this assumption can be made by comparing the pressures found from two pairs of wavelengths where some overlap exists. Thus, the pressures found from the pairs {5600 Å, 9900 Å} and {5600 Å, 6830 Å} agree quite closely. On the other hand the pressures found from pairs involving a shorter wavelength measurement seem significantly lower.

The pressure derived from the pair {3590 Å, 9900 Å} may be checked by employing the second method discussed above, since the observed polarization curve exhibits a positive extremum polarization near $\Phi = 90°$. The slight displacement of the peak towards shorter phase angle can be attributed to an increased polarization towards smaller phase angles from the cloud component, as indicated by the longer wavelength measurements. The second method, which makes use of the variation of the polarization with phase angle near 90°, makes lesser demands on the behavior of the cloud properties. From Eqs. (A6), (A11), and (A12) we obtain a cloudtop pressure which agrees within 15% with the {3590 Å, 9900 Å} pressure given in Table I.

The polarization measurements used in the calculations leading to Table I were those of Gehrels and Samuelson (1961); at $\lambda > 4000$ Å, approximately, those data agree well with more recent photoelectric polarimetry of Venus performed by Coffeen and Gehrels (1969). In the ultraviolet, near 90° phase angle, there is, however, strong evidence for time-variations in polarization. These variations are interpreted as a variation by a factor of 2 in the upper cloud pressures, with the value given in Table I being close to the maximum observed pressure. Using in part the formulas derived in the present paper and privately communicated to him, Coffeen (1969) has derived an upper cloud pressure of 55 mb from his ultraviolet polarization measurements. The difference between this result and our own, we believe, is due to an error by a factor of 2 in Coffeen's estimate of $F(\Phi = 0°)/F(\Phi = 90°)$.

The results of Table I may be compared with a value of 60 mb obtained by Dollfus (1966) from a photographic study of the extension of the cusps, and a value of about 10 mb obtained by Goody (1967) from a combination of transit and spectrometric measurements. Note that these results pertain to a level where the optical depth is very small, whereas our estimates pertain to a level close to optical depth unity. Earlier polarimetric observations by Dollfus (1957) led to an upper cloud pressure of roughly 90 mb, based on visual wavelengths.

In view of the good agreement of the cloudtop pressures found by our two methods for the ultraviolet wavelength, and the expected reliability of the second method, an ultraviolet pressure near the cloudtop of 34 mb is very likely within a factor of 2 or so of the actual value. The Mariner 5 measurements (Kliore *et al.*, 1968) revealed a steep and approximately constant temperature lapse rate at pressures of 300 mb and higher, a very shallow temperature gradient between the 100 and 300 mb levels, and an approximately isothermal region at pressures less than 100 mb. Comparing this profile with the above polarimetric results, we find that the cloud level characterized by an optical depth of at least 1 and, more likely, 3 or 4, is located high in the stratosphere. This result is quite unexpected, since the cloud particles are presumably opaque at infrared wavelengths, which would result in a significant temperature gradient near the 34-mb level and at higher pressure. In the Earth's atmosphere convective clouds rarely penetrate above the tropopause; only mother-of-pearl and noctilucent clouds are regularly found in the stratosphere. Conceivably the average particle diameter in the Cytherean stratosphere is sufficiently small ($<1\ \mu$) that the infrared cross section is much less than the visual cross section, and the optical depth of the particles in the stratosphere is small. Alternatively, there may be a second cloud region located high in the stratosphere. In this case the derived ultraviolet pressures "near the cloudtops" refer to a region somewhere between the high-altitude cloud layer and the principal cloud layer.

The polarimetric pressures derived from measurements at longer wavelengths are more questionable in view of the assumptions used to derive them, but it is worth noting that the pressures derived from the two long-wavelength pairs agree quite well. Let us assume that these pressures of about 130 mb are meaningful. These pressures lie within the transition region from steep to zero temperature gradient; such a location is where one might expect the level near optical depth unity to be located, in view of the probable effect of the cloud particles on the temperature gradient. The fact that higher pressures are found at visual than at ultraviolet wavelengths is most interesting. Venus reflects about twice as much light in the visual than in the ultraviolet. However, if we assume there is only one cloud layer, the single scattering albedo of the cloud particles would be quite high even at ultraviolet wavelengths and so little change would be expected in cloudtop pressure. However, it may also be possible to explain the large wavelength variation in polarimetric pressure again in terms of a high-altitude cloud which has a modest optical depth, and where particles have significantly smaller single-scattering albedos in the ultraviolet, leading to an enhanced relative return from this upper cloud layer. Such an explanation might also explain the patchy structure of the clouds as seen in the ultraviolet, a structure which is absent in the visual.

Temperatures Near the Cloudtops

The Mariner 5 spacecraft made an accurate determination of the scale height of the (approximately) isothermal portion of the atmosphere. When combined with temperature estimates this scale height yields the CO_2 mixing ratio, assuming the remainder of the atmosphere is N_2 or is otherwise fixed. We now attempt to estimate the range of temperatures near the cloudtops permitted by Earth-based observations.

To estimate the "cloudtop" or isothermal temperature we combine limb-darkening observations with absolute temperature measurements in the infrared. Sinton (1963) has found the disk-averaged brightness temperature of Venus to be $226° \pm 10°$K. Where μ is the angle between the local normal and the line of sight, the observed radiation flux, I, varies as $\mu^{1/2}$, except close to the limb (Pollack and Sagan, 1965*a*). Since the brightness temperature, T_B scales as $I^{1/n}$, where $n \simeq 6.3$ (Pollack and Sagan, 1965*a*) at these wavelengths and temperatures, the brightness temperature at the center of the disk is $233° \pm 10°$K. Sinton and Strong (1960) find only about 10°K variation in average brightness temperature throughout the 8–13 μ interval and a similar difference between the 8–13 μ region and the 3.75 μ region (Sinton, 1963). As some of this difference can be ascribed to a varying depth of penetration, the emissivity correction is probably at most 10°K, and so the thermometric temperature at the center of the disk is probably $233\ ^{+20}_{-10}$ °K. Westphal (1966) has observed close enough to the limb to obtain the asymptote in the limb-darkening behavior of I; his results imply that the asymptotic brightness temperatures very close to the limb, and hence the temperatures near the cloudtops and in the isothermal region, are 15K° below the temperature at the center of the disk. The limb darkening is almost entirely due to variations in the thermometric temperature with depth, since thermometric cloud temperatures significantly higher than indicated above are required before emissivity limb-darkening effects become important (Pollack and

Sagan, 1965*a*). Thus the cloudtop temperature is 218^{+20}_{-10} °K.

If we now combine the above temperature estimates with the scale height estimates of Mariner 5, and assume CO_2 and N_2 to be the major constituents of the atmosphere, we find a CO_2 mixing ratio of between 45 and 95%. In our estimates we have allowed for a 5% uncertainty in the value of the scale height. The above mixing ratios are compatible with the results of the Venera 4 chemical experiments of $90 \pm 10\%$ (Vinogradov and Surkov, 1968).

We note the the 15 K° difference between the temperature observed at the center of the disk and the extreme limb corresponds rather neatly to the temperature difference found by Mariner 5 between the isothermal region and the level at which the temperature gradient becomes steep. Thus the 8–13 μ emission originates mostly within this temperature transition zone of the atmosphere. This coincidence in temperature difference may reflect a much more diffuse distribution of aerosols in the region of small temperature gradient than in the region of strong temperature gradient.

Line Formation in the Venus Atmosphere

Gaseous absorption will take place both above and within the cloud layer. If the cloud is sufficiently diffuse, multiple scattering of solar photons can result in a much greater absorption within the clouds than above. Chamberlain and Kuiper (1956) pointed out that the observed decline of absorption line equivalent widths towards inferior conjunction apparently implies that line formation takes place mainly within the cloud layer and that atmospheric mixing ratios inferred on the basis of a simple reflecting layer model from the observed absorption would lead to overestimates of abundances. Below we attempt to explore further the validity of the multiple scattering model, by determining to what extent the simple reflecting layer model overestimates gaseous mixing ratios. The estimates of pressures near the cloudtops given in a previous section enable us to perform this additional test.

We now wish to estimate the base pressure implied by CO_2 absorption spectrometry on the basis of a simple reflecting layer model and compare the results with the previously obtained values of the pressures near an aerosol optical depth of unity. All the observations discussed here pertain to phase angles close to 50°. From an analysis of CO_2 absorption lines near 1.75 μ, which were on or close to the linear portion of the curve of growth, Connes *et al.* (1967) find ηw to be 3.3 km-atm. Here w is the amount of CO_2 in a vertical path down to some base level, and η is the ratio of the average pathlength of a photon to the vertical pathlength from the top of the atmosphere to the base level. If we take η to be 4, to allow for the slant path but not for multiple scattering, w will be 0.825 km-atm. This value agrees satisfactorily well with a much more poorly determined value for w of 2 km-atm, determined at 7820 Å (Spinrad, 1962). If we now take the CO_2 mixing ratio as 100%, we find the base pressure, p, to be 140 mb. [From hydrostatic equilibrium for pure CO_2, we find $P(\text{mb}) \simeq 170\ w$ (km-atm).] Using other values for the CO_2 mixing ratio compatible with the spacecraft measurements raises the derived base pressure somewhat.

In a similar fashion we deduce base pressures from bands that are on the square root and logarithmic portions of the curve of growth. Planimetering published spectra obtained by Kuiper (1962), we find an equivalent width of 52 cm^{-1} for the band complexes between 1.20 and 1.25 μ, and 223 cm^{-1} for the complex between 1.515 and 1.67 μ. We next apply the empirical formulas of HBW (Howard, Burch, and Williams, 1956) for these wavelength regions, with the constants given by Pollack (1969). The pressure in the HBW formulas was taken as half the base pressure, in accordance with the Curtis-Godson approximation, and was augmented by a factor of 1.3 to allow for self-broadening of CO_2. A value of $\eta = 4$, typical in the simple reflecting model, was adopted. Finally the above relationship between P and w was employed. The base pressure was found to be 170 mb for the 1.22-μ complex and 220 mb for the 1.6-μ complex. We note that the 1.6-μ complex is on the logarithmic part of the curve of growth. In a previous analysis of the region, Kaplan (1961) assumed this complex was on the square root portion. The above pressure estimate may be too high; the spectral regions in question encompass several bands. The influence of some of the weaker bands may be greater for the long path of the Venus atmosphere than inferred from the HBW data for much shorter paths. Accordingly we repeated the calculations with equivalent widths of 35 and 150 cm^{-1} which represent the contribution of the strongest bands on Venus, and find base pressures of 110 and 95 mb. The true base pressure should lie between 110 and 170 mb for the 1.22-μ complex.

We now compare the above base pressures with the cloud pressure of 130 mb obtained from the longer wavelength polarization data and a value between 100 and 300 mb implied by location of the transition temperature gradient zone in the Mariner 5 results. The spectroscopically derived base pressures are quite close to the pressures near optical depth unity. This would imply that much of the line formation for CO_2 takes place above the clouds and that, to first order, mixing ratios inferred from a simple reflecting layer model (at least at moderate phase angles) are correct. A similar conclusion was reached by Gray (1968) by comparing space vehicle with a smaller range of spectroscopically determined pressures. However, if we were to use the ultraviolet cloudtop pressures of 35 mb, then the predictions of the simple reflecting layer model would be about a factor of 5 too small.

It is quite interesting that the base pressures inferred from lines on the linear portion of the curve of growth agree with those on the square root and logarithmic portions. It is not clear that such a result is compatible with the pure multiple scattering model, since the fractional absorption varies as the square root of the line strength even for very weak bands for this model, provided the single scattering albedo of cloud particles can be taken as unity. As a result the multiple scattering model might lead one to expect that base pressures inferred by a simple reflecting layer model for lines on the linear portions of the curve of growth would be much larger than base pressures inferred from stronger lines.

We now return to a consideration of the phase dependence of the equivalent width of absorption features. To first approximation the equivalent width of the 8689-Å band varies as μ, except close to inferior conjunction (Chamberlain and Kuiper, 1956), whereas a simple reflecting layer model would predict a μ^{-1} dependence for lines in the linear and square root domains. The values of base pressure given above are disk-averaged values. At the center of the disk the base pressure will be three times that of the disk-averaged value, or 400 mb. Near $\mu = \frac{1}{4}$ the equivalent width appears to approach an asymptote. The asymptotic base pressure is then about $\frac{1}{16}$ the value of the disk center, or about 25 mb.

The value of the base pressure at the center of the disk represents an upper limit to the pressure level of the atmosphere involved in CO_2 line formation. The base pressure at the center of the disk is the largest value of any portion of the disk and it is an overestimate in neglecting the effect of multiple scattering, which enhances η and so lowers p. It would therefore appear that almost all line formation takes place within the transition zone and isothermal region of the atmosphere. The average temperature for the line formation will lie between the isothermal value and a value 15 K° higher. The average temperature of line formation then is predicted to lie between 208° and 253°K. Temperatures of 200–250°K inferred by Gray and Schorn (1968) and by Spinrad (1968) are compatible with these estimates, while higher temperatures have been obtained by Belton and Hunten (1968). These higher values may in part be the result of observational errors, and in part of the invalidity of using a homogeneous line formation model, i.e., one in which all lines are formed at the same effective temperature.

We are now in a position to assess the implications of the observed water vapor lines for the hypothesis that the clouds are composed of water. Hunten, Belton, and Goody (1968) have argued that the strength of the water vapor line they detected is some three orders of magnitude weaker than expected for line formation within a water cloud. Our analysis above indicates that this conclusion is premature. Hunten, Belton, and Goody employed a homogeneous scattering model with a temperature of 270°K, and all the line formation taking place by multiple scattering in the saturated cloud region. Since the water vapor saturation curve is very temperature-sensitive, the effect of going to the lower temperatures derived above is to remove most of this discrepancy. At a temperature of about 215°K the use of a purely multiple scattering model would produce no discrepancy. Furthermore, it is clear that a significant portion of the line formation process occurs in the stratosphere, where conditions may be far from saturation.

We now employ a simple reflecting layer model to obtain a first approximation to the amount of water vapor which should be detected if the Venus clouds are ice. We recall that the simple reflecting layer model yields reasonable results for CO_2. If water vapor has a uniform mixing ratio above some level in the Venus atmosphere, the number of gm cm^{-2} of water vapor, w_{H_2O}, will equal 0.61 P_{H_2O}, when P_{H_2O} is the partial pressure of water vapor at the level of interest in units of mm Hg. The water vapor in the Earth's stratosphere is at 1% saturation; if we assume this to be true for Venus as well, we find between 2.5×10^{-5} gm cm^{-2} (0.25 μ) and 10^{-3} gm cm^{-2} (10 μ) in the Venus stratosphere, depending on the choice of stratospheric temperature. If, on the other hand, we assume 100% saturation, we find the water content to vary between 2.5×10^{-3} gm cm^{-2} (25 μ) and 10^{-1} gm cm^{-2} (1000 μ). Similarly, the amount of water vapor in the transition zone, where saturation is assumed to hold, will vary from about 6×10^{-3} gm cm^{-2} (60 μ) to 2×10^{-1} gm cm^{-2} (2000 μ). We have seen above that most of the stratosphere seems involved in the line formation process, and only a part or none of the transition region is involved. Estimates of observed amounts of water vapor range from 10^{-4} gm cm^{-2} (1 μ) to 10^{-2} gm cm^{-2} (100 μ) (Belton and Hunten, 1966; Kuiper, 1968; Schorn *et al.*, 1969). These amounts are compatible with the amounts expected if an ice cloud is present, within the observational uncertainties in detected amounts of water vapor, in cloudtop temperature, and in the fraction of the transition zone involved in the line formation process. Other arguments for and against Venus ice clouds have been published (e.g., Rea and O'Leary, 1968; Pollack and Sagan, 1968; Coffeen, 1969); we wish here only to point out that there is no compelling argument from spectroscopic searches for water vapor against such clouds.

We next inquire into ways of understanding the phase dependence of the equivalent width. There seem to be several prospects for understanding the observed weakness of lines towards inferior conjunction, i.e. towards higher values of μ^{-1}, without invoking a multiple scattering model. We might postulate that the clouds have greatly varying altitudes from topographical position to topographical position. As we go towards inferior conjunction our line of sight tends preferentially to intercept the higher lying clouds. The greatly irregular shape of the terminator implies a large enough variation in cloud altitudes (several tens of kilometers[3]) for this effect to override the usual μ^{-1} effect. Alternatively, we showed in the discussion of the ultraviolet polarization curves that there may be a very high-lying aerosol layer which contributes to the light seen at ultraviolet frequencies. Some contribution would be expected at longer wavelengths; and possibly some of the phase effect is due to the larger contribution to the observed radiation from the high-level aerosol

[3] Consider Venus at quadrature, where the radius is 10″ of arc. To see an irregularity implies it is ~1 sec of arc or 600 km in horizontal extent. For this to be illuminated by the Sun it must be at an altitude $(600)^2/2(6000) = 30$ km higher than the adjacent clouds.

layer near inferior injunction because of the increase in the slant path optical depth of this layer toward inferior conjunction.

Finally, we caution theoreticians working on the line formation problem: Cloud properties inferred from the observed lines will pertain to the transition region and perhaps the stratosphere as well; accordingly, they are inapplicable to the zone below the transition region. In particular, if the cloud is found to be diffuse from spectral observations, such information only provides a lower limit to the aerosol density below the transition zone. Similarly one would not expect *a priori* that the mixing ratio of cloud particles to gas molecules is constant within the transition zone or in the stratosphere.

We now summarize the main points of this section. The use of a simple reflecting layer model leads to predicted base pressures for the CO_2 near-infrared absorption lines that are either quite close to or at most a factor of 5 in excess of the "observed" value. Thus the simple reflecting model may be used for crude analysis of absorption features. Line formation occurs within the stratosphere and perhaps within part of the transition zone. Accordingly, temperatures of between 208° and 253°K should characterize the absorption features. Furthermore, within the present uncertainties, the predicted strengths for the water vapor lines expected if an ice cloud is present on Venus are compatible with the observations. A multiple scattering model is not the only one capable of explaining the phase dependence of the absorption line strength. Either a single cloud layer with individual clouds having a large range of altitudes or a model with two cloud layers might account for the phase observations.

An Analysis and Comparison of the Mariner 5 and Venera 4 Temperature–Pressure Measurements

We wish to examine the consistency and implications of the recent spacecraft measurements on the structure of the Venus atmosphere. Figure 1 compares a portion of the Venera 4 pressure and temperature measurements (Avduevsky *et al.*, 1968) with corresponding (revised) determinations by Mariner 5 (Kliore *et al.*, 1968). The Mariner 5 observations were obtained only to the 5-atm level and have been adiabatically extrapolated to higher pressures. We have scaled the Mariner 5 data appropriately to obtain the indicated curves for carbon dioxide mixing ratios of 50, 85, and 99%. The remainder of the atmosphere is assumed to be composed of nitrogen. The Venera 4 measurements at selected points, together with their estimated errors, are indicated by crosses. The two data sets appear consistent and superficially a mixing ratio of about 85% would seem to be indicated. However, in making this comparison we must remember that the Venera 4 points refer to a region quite close to the nominal equator of Venus, while the Mariner 5 data points refer to a latitude of about 37°. Interferometric observations at 10 cm by Clark and Kuz'min (1965) have indicated that the temperature difference between the equator and pole is about 25% of the mean surface temperature. It is difficult directly to apply these data to obtain the surface temperature difference between the Mariner 5 and Venera 4 locales, since we do not know the functional dependence of surface temperature on latitude. However, temperature differences of up to 75 K° would not seem unreasonable (see Pollack and Sagan, 1965*b*). If we assume that the temperature structure of the atmosphere is the same at the two locales, then at a given pressure level the temperature at the Mariner 5 locale will be approximately $\bar{T}\Delta T_s/\bar{T}_s$ cooler, where $\bar{T}$ denotes an average temperature between the two positions, $\bar{T}_s$ an average surface temperature, and ΔT_s the difference in surface temperature. This means that the Venera 4 data points should be moved horizontally and to the left in Fig. 1 by as much as 40 K°. For the maximum displacement a carbon dioxide mixing ratio of about 50% is indicated. We conclude that the carbon dioxide mixing ratio is between 50 and 85%.

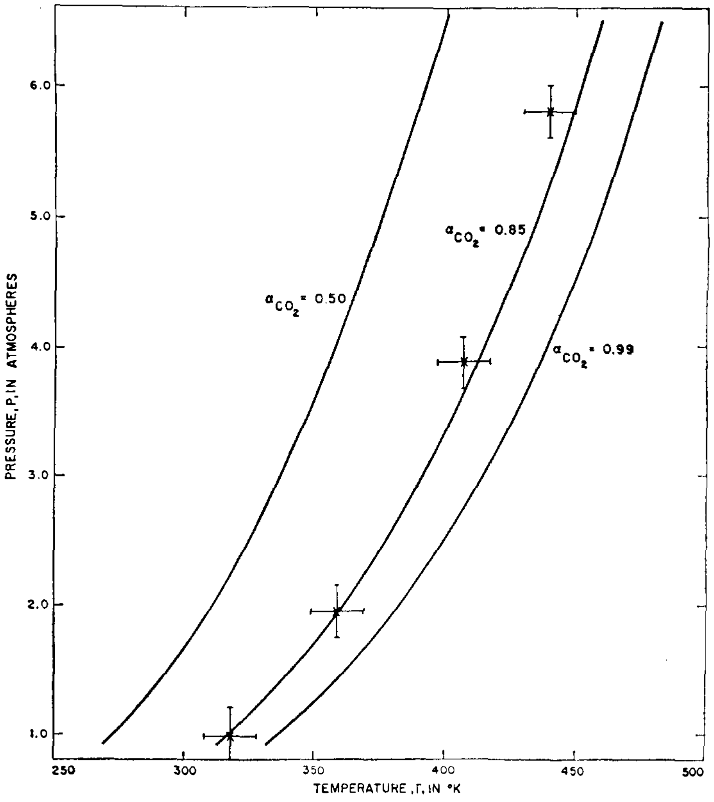

FIG. 1. A comparison between the determination of the atmospheric structure of Venus by Mariner 5 (solid curves) and Venera 4 (crosses). The carbon dioxide mixing ratio used in converting the Mariner 5 results to temperature and pressure is indicated. The extent of the crosses corresponds to the error estimates given by the Venera 4 experimenters.

An interesting implication of this result is that there must be some other major constituent of the Venus atmosphere besides carbon dioxide. Perhaps the most reasonable candidate is nitrogen. One of the chemical analysis experiments aboard the Venera 4 spacecraft indicated an upper limit to the N_2 mixing ratio of 7%. However, since the experiment involved the potential breakage of a heated wire, it is unclear how certain this upper limit is. We recall that a similar experiment gave a positive measurement of oxygen that is several orders of magnitude larger than the spectroscopically determined upper limits. Until more details on the calibration of the nitrogen detection experiment are available, we believe that the 7% upper limit should not be taken too literally.

As a second application of the above limits on the carbon dioxide mixing ratio, we consider the ability of water condensation clouds to form. The chemical experiments performed by Venera 4, not based on the breakage of a heated wire, indicated that the water vapor mixing ratio lies between 0.1 and 0.7% in the lower atmosphere. Saturation would be achieved between the 240° and 270°K level (Pollack and Sagan, 1968). Since the stratospheric temperatures implied by the Mariner 5 scale height (Kliore *et al.*, 1968) and by the above limits on the CO_2 mixing ratio are between 210° and 237°K, it appears that ice condensation clouds would form.

We next consider the temperature structure of the atmosphere at levels exhibiting a steep temperature gradient

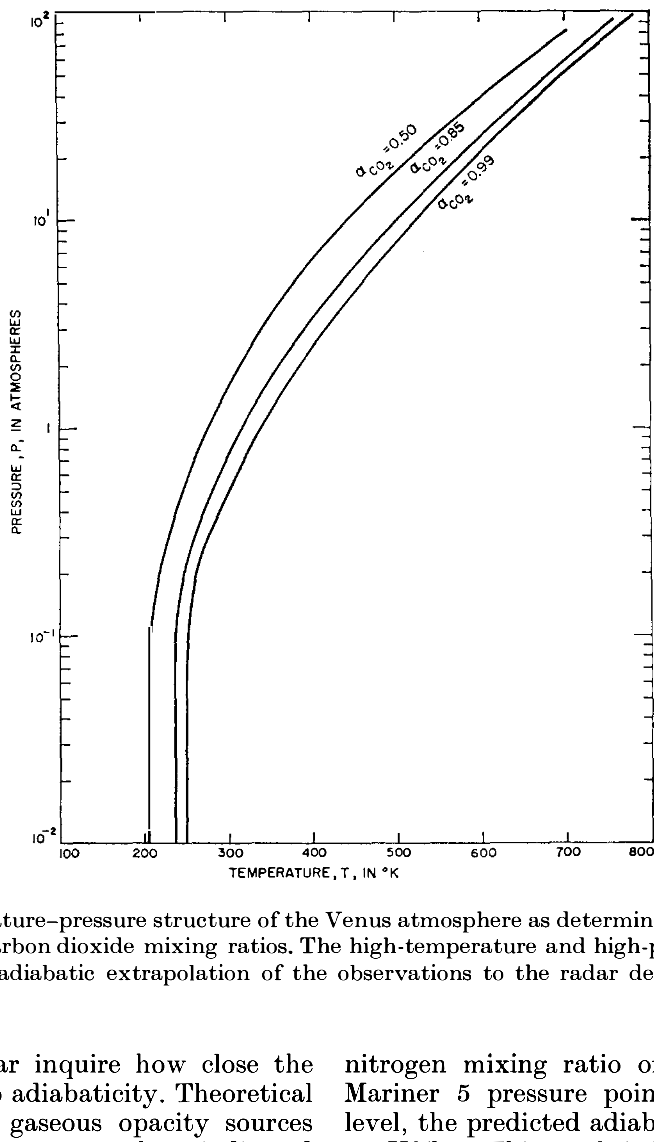

FIG. 2. Temperature–pressure structure of the Venus atmosphere as determined by Mariner 5 for various assumed carbon dioxide mixing ratios. The high-temperature and high-pressure end points correspond to an adiabatic extrapolation of the observations to the radar determined radius of 6050 km.

and in particular inquire how close the atmosphere is to adiabaticity. Theoretical calculations for gaseous opacity sources (Pollack, 1969; Sagan, 1969) have indicated that the lower portion of the atmosphere should be to first approximation in convective equilibrium. Venera 4 measured an average temperature gradient of 8.9 K°/km between the 400° and 543°K levels of the atmosphere. Richard Wattson has kindly supplied us with outputs of an adiabatic temperature gradient computer program that allows for the variation of specific heat with pressure and temperatures and employs the van der Waals equation of state (see Wood, Wattson, and Pollack, 1968). For a CO_2 mixing ratio of 85%, a nitrogen mixing ratio of 15%, and the Mariner 5 pressure point at the 400°K level, the predicted adiabatic lapse rate is 8.9 K°/km. This result is insensitive to the exact choice of mixing ratios. Thus the region of the atmosphere at temperatures in excess of 400°K is quite close to adiabatic.

On the other hand, between the 300° and 400°K levels, the Venera 4 data indicate a lapse rate of 8.1 K°/km compared to a predicted value of 9.5 K°/km. The reality of this discrepancy is indicated by the close agreement of the Venera 4 results with the Mariner 5 data. For a CO_2 mixing ratio of 85%, a lapse rate of 8.3 K°/km is found between the above temperature levels,

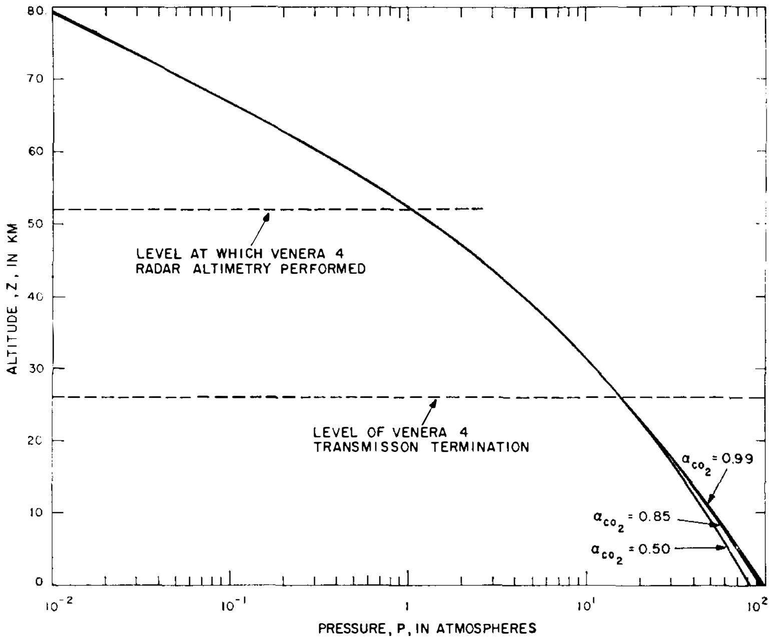

FIG. 3. Pressure as a function of altitude in the Venus atmosphere for various assumed carbon dioxide mixing ratios. The Mariner 5 results have been adiabatically extrapolated to the radar determined radius of 6050 km.

again with only small changes between the 50 and 100% mixing ratio cases. The Mariner 5 results indicate somewhat smaller temperature gradients pertain from the 300°K level to the transition zone where very small gradients are encountered. With water vapor mixing ratios such as were apparently found by Venera 4, a water cloud can be expected between the tropopause and the 240° to 270°K level (see Pollack and Sagan, 1968); temperature gradients compatible with the observations would be expected in this region due to the release of latent heat of condendastion (*ibid.*). Even in this case it would seem that the temperature gradient is slightly subadiabatic between the cloud bottom and the 400°K level. As a working model of the Venus atmosphere we have used an adiabatic lapse rate for temperatures of 400°K and higher (or pressures in excess of 4 atm) and have subdivided the region between the tropopause and the 400°K level (the 378°K level for the 50% CO_2 case) into five zones of constant lapse rate. The resultant atmospheric structure is shown in Figs. 2, 3, and 4, where we have considered the surface to be located at the radar radius of 6050 km (Ash *et al.*, 1968), and have used the Mariner 5 data to define the model. As there is little expected diurnal temperature variation and as the Mariner results refer to a midlatitude region, the structure given below may be taken as a mean structure for the planet.

The use of the Mariner 5 estimates of distance from the center of the planet and the radar radius to define the location of the surface leads to an average surface temperature and pressure of about 750° K and 90 atm as shown in Figs. 2, 3, and 4. The above temperature estimates are incompatible with the claim that the last-measured Venera 4 point of 544° K was at or near the surface. A 750° K surface is in good agreement with estimates from passive microwave radiometry (Sagan, 1962), and with surface temperature estimates made from the radar radius

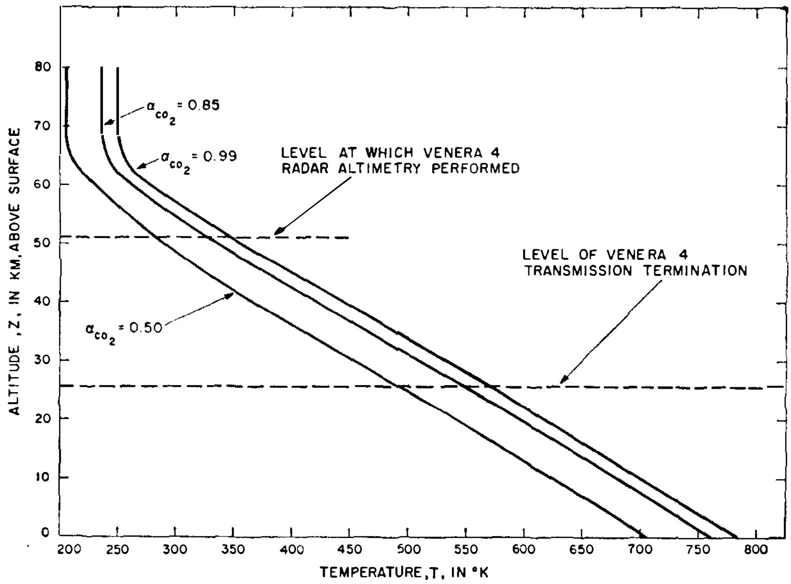

FIG. 4. Temperature as a function of altitude in the Venus atmosphere for various assumed carbon dioxide mixing ratios. The Mariner 5 results have been adiabatically extrapolated to a radar determined radius of 6050 km.

(Sagan, 1967) before the Mariner 5 and Venera 4 encounters. Furthermore, Wood, Wattson, and Pollack (1968) show that ground-based radio brightness temperature and radar cross-section measurements distinctly favor this high-temperature, high-pressure case. Except possibly for a narrow boundary region near the surface, the atmosphere will not show any appreciable diurnal variation, as mentioned above; accordingly, the low value of the surface temperatures claimed by Avduevsky *et al.* (1968) cannot be explained by attributing them to the night side of the planet. The proximity of the Venera 4 measurements to the equator precludes a reconciliation in terms of a latitudinal temperature gradient. Finally, it seems quite unlikely that Venera 4 fortuitously landed on a highland some 25 km or more above the mean surface. This contention is supported by radar data indicating significantly smaller elevation differences (Ash *et al.*, 1968), and by the clear difficulty of maintaining such elevation differences with the tensile and yield strengths of geochemically abundant materials in the Venus gravity field.

Let us now examine the support for the contention (Avduevsky *et al.*,1968) that the last Venera 4 measurement refers to the surface. The Venera 4 capsule employed a radar altimeter designed to indicate when the 26-km level above the surface was reached. However, many radar altimeters are to a certain degree ambiguous. For example, with pulsed systems, distances which require times in excess of the time between pulses can be misinterpreted as smaller distances. We need far more information about the design of the radar altimeter to determine whether such ambiguities are important. We emphasize that only one radar altimeter measurement was made. Also the altimeter detection of the 26-km altitude occurred quite close to parachute deployment; possibly such deployment could have adversely effected the altimeter. While the transmitter ceased sending data quite close to a point about 26 km below the level at which the radar altimeter gave its reading, we note that the total travel time, covering 28 km, of 93 min was quite close to the anticipated lifetime of the battery used for power, about 100 min. Studies of aerodynamic drag on the Venera 4 capsule (Avduevsky *et al.*, 1968; Reese and Swan, 1968) confirm that transmission ceased 26 km below the point determined by radar to be at 26-km altitude; but this in fact only confirms a 26-km altitude difference between radar measurement and the end of transmission—it does not bear directly on the question of the distance to the surface.

A very interesting possibility emerges from Fig. 4. Considering, for convenience, the 85% CO_2 mixing ratio case, we see that 550° K point, lying 26 km below the altitude at which the radar altimetry was performed, was itself at an altitude almost exactly 26 km above the 750°K surface. Thus, we raise the possibility that the Venera 4 radar altimeter—about which very little detailed information has been published to date—was ambiguous to all integral multiples of 26 km. If this is the case, the point reported as 26 km above the surface was actually $2 \times 26 = 52$ km high. [Added note: This identical point has been made independently by Eshleman, *et al.* (1968), in the November 8 issue of *Science*.] But regardless of the details of the radar altimetry, there seems to be serious reason to doubt that the 550°K, 20-atm point refers to the surface of Venus.

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Appendix

Relationship Between Polarization Measurements and Cloudtop Pressures

We wish to obtain several expressions for pressures near the cloudtops in terms of the observed polarization of the entire illuminated disk. We begin by formally separating the measured photons into (1) ones that have been scattered by the atmospheric gases, but not by the cloud particles, and (2) ones that have been scattered by cloud particles. The latter may or may not be scattered also by the atmospheric gases. The first component we denote by the subscript a; in many cases this component will represent only a small fraction of the total number of photons received, but because of its high polarization its influence may be significant. The second component will in general have a low polarization because of multiple scattering within the clouds. For this reason we include within this group photons that have been scattered by atmospheric gases before or after multiple scattering by the clouds, prior to escape into space. Thus the observed polarization, p, may be written as

$$p = \frac{p_a B_a + [B_T - B_a] p_c}{B_T}, \qquad \text{(A1)}$$

where the subscript a refers to the first (atmospheric) component, the subscript c refers to the second (cloud) component, the subscript T represents the total number of photons, and B is the brightness of the entire illuminated disk.

We now suppose that the Rayleigh scattering optical depth is small, so that only single scattering events need be considered for the first component. In this case $B_a \ll 1$ and $p_a \simeq 1$. We further suppose that $B_T \sim 1$ and $p_c \ll 1$, in accord with the observation that $p \ll 1$. Under these assumptions Eq. (A1) simplifies to

$$p = p_a(B_a/B_T) + p_c. \qquad \text{(A2)}$$

There are two methods which we will consider for obtaining the ratio B_a/B_T. The first method involves the assumption that p_c is independent of wavelength at a given phase angle. If this is true we may obtain B_a/B_T simply by subtracting polarization measurements made at two different wavelengths. Noting that for small Rayleigh scattering optical depths, τ, $B_a(\lambda) \propto \tau \propto \lambda^{-4}$, where λ is the wavelength, and that p_a is wavelength-independent, we obtain from Eq. (A2) the following expression for the difference in the value of the polarization measured at the wavelengths λ_1 and λ_2:

$$p(\lambda_1) - p(\lambda_2) = p_a \frac{B_a(\lambda_1)}{B_T(\lambda_1)}\left[1 - \left(\frac{\lambda_1}{\lambda_2}\right)^4 \frac{B_T(\lambda_1)}{B_T(\lambda_2)}\right], \quad \text{(A3)}$$

where $\lambda_1 < \lambda_2$. Note that generally $(\lambda_1/\lambda_2)^4 \ll 1$ so that the exact value of $B_T(\lambda_1)/B_T(\lambda_2)$ does not significantly affect the derived value of $B_a(\lambda_1)/B_T(\lambda_1)$. It is best to apply Eq. (A3) to observations made at phase angle 90°, where p_a achieves its maximum value.

As a result of molecular asymmetry p_a is not exactly 1.00 at phase angle $\Phi = 90°$. Rather p_a will be given by (van de Hulst, 1952)

$$p_a(\Phi = 90°) = \frac{3 - f}{3f - 1}, \quad \text{(A4)}$$

where f is the molecular asymmetry factor. For pure CO_2, $p_a(\Phi = 90°)$ is 0.82. The more general expression for p_a at any phase angle is:

$$p_a = \frac{1 - [\cos^2\Phi + \frac{1}{2}(f-1)\sin^2\Phi]}{f + \cos^2\Phi + \frac{1}{2}(f-1)\sin^2\Phi}. \quad \text{(A5)}$$

When the atmospheric component is sufficiently important to produce a noticeable peaking in the observed polarization curve near phase angle 90°, a second, more accurate, method of extracting B_a/B_T may be used. We assume that both p_c and B_a/B_T vary linearly with phase angle between $90° - \theta$ and $90° + \theta$. We now make use of Eqs. (A2) and (A5) and subtract the observed polarization of 90° from the average values of the observed polarization at $90° - \theta$ and $90° + \theta$. The resulting equation for B_a/B_T is

$$p(\Phi = 90°) - \tfrac{1}{2}[p(\Phi = 90° - \theta) + p(\Phi = 90° + \theta)] = \left(\frac{B_a}{B_T}\right)_{\Phi=90°} [p_a(\Phi = 90°) - p_a(\Phi = 90° - \theta)]. \quad \text{(A6)}$$

Note that according to equation (A5), $p_a(\Phi = 90 - \theta°)$ equals $p_a(\Phi = 90 + \theta°)$.

The ratio B_a/B_T at a phase angle of 90° may be obtained from Eqs. (A3) or (A6). We now proceed to relate B_a/B_T to the Rayleigh scattering optical depth τ and the photometric observations, and then relate τ to the pressures near the cloudtops. We recall that B_a and B_T refer to the brightness of the entire illuminated disk. Consider a small area on the disk where the local normal makes an angle arccos μ with the line of sight and with an azimuthal angle ϕ. For an optically thin Rayleigh scattering layer the brightness of this area, $B_a(\mu,\phi)$, or equivalently its specific intensity, is given by (Pollack, 1967):

$$B_a(\mu, \phi) = \frac{\mathscr{F}_{\odot}\tau}{4\pi\mu} \frac{f + \cos^2\Phi + \frac{1}{2}(f-1)\sin^2\Phi}{(4/3)f}, \quad \text{(A7)}$$

where $\mathscr{F}_{\odot}$ is the solar flux. In the above equation we have corrected the Rayleigh scattering phase function used by Pollack (1967) to allow for molecular assymmetry (see van de Hulst, 1952). B_a is simply the flux integral of $B_a(\mu,\phi)$. The integration is carried out only over the illuminated portion of the planet. At a phase angle of 90°, the integration is readily carried out as μ ranges from 0 to 1 and ϕ from 0 to π. We find

$$B_a = \int_0^{\pi}\int_0^1 B_a(\mu, \phi)\,\mu\, d\mu\, d\phi = \pi B_a(\mu = 1), \quad \Phi = 90°. \quad \text{(A8)}$$

Finally, the total brightness B_T may be expressed in terms of the diffusion factor, K, which is the ratio of the light received from a planet at a given phase angle to that which would be received from a white, Lambert scattering disk of the same area, which is oriented normal to the Earth and Sun. $B_T(\mu,\phi)$ is simply $K(\mu,\phi)\mathscr{F}_{\odot}/\pi$ (Pollack, 1967) and so B_T will be given by

$$B_T = \bar{K}\tfrac{1}{2}\mathscr{F}_{\odot}, \quad \text{(A9)}$$

where $\bar{K}$ represents the appropriate average value of $K(\mu,\phi)$. In turn $\bar{K}$ is straightforwardly related to the geometric albedo, A_g, and the ratio of the observed flux of radiation F at phase angle 90° to that at 0°, both corrected to standard distances

$$\bar{K} = \frac{2F(\Phi = 90°)}{F(\Phi = 0°)} A_g. \quad \text{(A10)}$$

The factor 2 allows for the fact that the illuminated area of the planet is twice as large at phase angle 0° as at 90°. The flux ratio and geometric albedo are both commonly observed quantities. Combining Eqs. (A7) through (A10), we obtain an expression for B_a/B_T at phase angle 90° in terms of the Rayleigh scattering optical depth and observed photometric quantities

$$\frac{B_a}{B_T} = \frac{(\frac{3}{2}f - \frac{1}{2})}{\frac{16}{3}f} \frac{F(\Phi = 0°)}{F(\Phi = 90°)} \frac{\tau}{A_g}. \quad \text{(A11)}$$

Finally the Rayleigh scattering optical depth, τ, may be related to the pressure at the base of the scattering region, P. Here this pressure is that near the cloudtops. P and τ are related (Pollack, 1967) by

$$P = \frac{\rho'}{\beta'} g\tau \left[\frac{m_0}{m'} \frac{f'(n'-1)^2}{\sum_i f_i v_i (n_i - 1)^2}\right], \quad \text{(A12)}$$

where ρ' is the density of air at STP, 1.29×10^{-3} gm/cm^2; g is the acceleration of gravity, 872 cm sec^{-2} for Venus; β' is a scattering parameter at STP for air, given by van de Hulst (1952) at various wavelengths; m_0 and m' are the mean molecular weights of the planetary atmosphere and of air, respectively; n' the index of refraction of air at STP; n_i is the refraction index of the ith constituent of the planetary atmosphere; and v_i the fraction by volume of constituent i in the planetary atmosphere. The quantities f', n', f_i, and n_i are given by van de Hulst (1952). Note that the units of P are dynes cm^{-2} as all the above quantities are in cgs units; thus P must be multiplied by 10^{-3} to obtain a result in units of mb. The above relationship between P and τ depends only upon the assumption of the validity of the perfect gas law and hydrostatic equilibrium, and is independent of the temperature structure of the atmosphere.

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