Venus: Vertical Transport Rates in the Visible Atmosphere[1]

RONALD G. PRINN

Dept. of Meteorology, Massachusetts Institute of Technology, Cambridge 02139

(Manuscript received 15 April 1974, in revised form 7 May 1974)

ABSTRACT

The cloud particles on Venus are sufficiently small for their vertical distribution to be strongly affected by atmospheric turbulence. Reasonably firm estimates of the vertical distribution of cloud particles can be made from current interpretations of refraction, polarization and absorption band data. These enable very firm upper limits to be placed on average vertical mass diffusion coefficients K in the visible atmosphere:

$$K<7\times 10^4 \text{ cm}^2\text{ sec}^{-1},\ 5\text{–}50 \text{ mb};\quad K<2\times 10^5 \text{ cm}^2\text{ sec}^{-1},\ 50\text{–}155 \text{ mb}.$$

Here K includes all scales of vertical motion. These upper limits are independent of the cloud particle composition or mode of formation, and of changes in the particle size distribution with altitude.

1. Introduction

Vertical transport rates are a very important input to studies of the aeronomy of planetary atmospheres. Unfortunately, it has proved very difficult to assess these rates using remote observations of the planets and widely different assumptions have consequently been made concerning them [*cf.* McElroy and Donahue (1972) and Parkinson and Hunten (1972) on Mars; McElroy *et al.* (1973) and Prinn (1973) on Venus]. In this paper we demonstrate that the visible cloud particles on Venus can be used as a convenient tracer for vertical mixing in this particular atmosphere. The subsequent limits placed on mixing rates using this procedure have important consequences for Venusian photochemistry and dynamics.

2. Transport considerations

The very rapid zonal (and possibly meridional) motions observed at and above the visible cloud level on Venus (Smith 1967; Traub and Carleton, 1971; Marov *et al.*, 1973) imply day-side to night-side horizontal transfer times for cloud particles of $\sim 10^5$ sec. If we assume that vertical transfer times for these particles are much longer (we will later verify this assertion), it is appropriate to define a long-term planetary average cloud particle distribution in which the effects of horizontal transports are averaged out and vertical transport only is considered.

With this assumption the steady-state mass of cloud material per unit volume (ρ_c), at altitude z, can be obtained by solving the two coupled mass continuity equations

$$\frac{\partial}{\partial z}\left[K\rho\frac{\partial}{\partial z}\left(\frac{\rho_c}{\rho}\right)+\langle w\rangle\rho_c\right]+\frac{\rho_g}{t_0}=0, \tag{1}$$

$$\frac{\partial}{\partial z}\left[K\rho\frac{\partial}{\partial z}\left(\frac{\rho_g}{\rho}\right)\right]-\frac{\rho_g}{t_0}=0. \tag{2}$$

Here ρ is the atmospheric mass density and t_0 a time constant for the thermodynamically irreversible production of cloud material from a precursor gas with mass density ρ_g by condensation or by thermochemical or photochemical reactions. For a simple dust cloud $\rho_g=0$ and $t_0=\infty$. We are referring here to the topside of the clouds so thermodynamically irreversible cloud mass loss by evaporation is not considered. Since particle coagulation does not result in loss of total cloud mass, explicit consideration of this phenomenon is avoided.

The parameter K in (1) and (2) is a vertical mass diffusion coefficient which will include all scales of vertical motion. If vertical wind velocities in the mean planetary circulations are sufficiently weak than K will essentially be the vertical eddy mass diffusion coefficient. The mass-weighted mean sedimentation velocity for cloud material, $\langle w\rangle$, used in (1) is defined by

$$\langle w\rangle=\frac{\left[\int_0^\infty (\tfrac{4}{3}\pi r^3 D)\left(\frac{2gD}{9\eta}r^2\right)f(r)dr\right]}{\left[\int_0^\infty (\tfrac{4}{3}\pi r^3 D)f(r)dr\right]},$$

$$=\frac{2gD}{9\eta}\bar{r}^2. \tag{3}$$

[1] Contribution No. 105 from the MIT Planetary Astronomy Laboratory.

Here D is the density of the cloud material, $f(r)$ the distribution function for cloud particle radii, g the gravitational acceleration, and η the absolute viscosity (of CO_2 in this case). We have assumed that the downward sedimentation velocity of individual particles of radius r is given by the Stokes-Cunningham formula for laminar flow which is valid for the slowly settling cloud particles we will consider. The "effective sedimentation radius" $\bar{r}$ is simply

$$\bar{r}=\left\{\left[\int_0^\infty r^5 f(r)dr\right]\Big/\left[\int_0^\infty r^3 f(r)dr\right]\right\}^{\frac{1}{2}},$$
$$=\langle r\rangle(1+3\delta+2\delta^2)^{\frac{1}{2}}. \tag{4}$$

Hansen and Arking (1971) and Hansen and Hovenier (1974) have analyzed measurements of polarized light reflected from Venus and have inferred a very firm estimate of $f(r)$, namely

$$f(r)=r^{(1/\delta)-3}\exp\left(-\frac{r}{\delta\langle r\rangle}\right), \tag{5}$$

where the "mean effective radius" for particle extinction $\langle r\rangle=1.05\pm0.1$ μm, and the "effective variance" $\delta=0.07\pm0.02$. The largest possible value for $\bar{r}$ is 1.3 μm and is obtained using (4) with the largest possible values of $\langle r\rangle$ and δ. The similarity in the values of $\langle r\rangle$ and $\bar{r}$ is due to the remarkably narrow range of particle sizes implied by the polarization measurements (i.e., small δ values).

Using the logical boundary conditions, ρ_c/ρ and $\rho_g/\rho\rightarrow0$ as $z\rightarrow\infty$, and given $\rho_c=\rho_c(0)$ and $\rho_g=\rho_g(0)$ at $z=0$, the appropriate solutions to (1) and (2) are

$$\left.\begin{aligned}
\rho_c&=\rho_c(0)\exp\left[-\left(\frac{\langle w\rangle}{K}+\frac{1}{H}\right)z\right]\\
&\quad+\frac{\rho_g(0)\left(\dfrac{1}{H_g}-\dfrac{1}{H}\right)}{\left(\dfrac{1}{H_g}-\dfrac{1}{H}-\dfrac{\langle w\rangle}{K}\right)}\times\left\{\exp\left[-\left(\frac{\langle w\rangle}{K}+\frac{1}{H}\right)z\right]-\exp\left[-\frac{z}{H_g}\right]\right\}\\
&\geqslant\rho_c(0)\exp\left[-\left(\frac{\langle w\rangle}{K}+\frac{1}{H}\right)z\right]\\
\rho_g&=\rho_g(0)\exp\left[-\frac{z}{H_g}\right]\\
\frac{1}{H_g}&=\frac{1}{2H}+\left(\frac{1}{4H^2}+\frac{1}{Kt_0}\right)^{\frac{1}{2}}\geqslant\frac{1}{H}
\end{aligned}\right\}. \tag{6}$$

Here H is the atmospheric mass scale height. In reality, t_0 will generally vary with altitude z (e.g., in a condensing cloud due to changes in temperature and condensation nuclei populations, or in a photochemical cloud due to extinction of dissociating photons). We will not discuss alternative t_0 values here although they can be accurately defined in specific cloud models for Venus (e.g., Prinn, 1973). We will be concerned here only with the lower limit to $\rho_c/\rho_c(0)$ expressed in (6). This implies

$$K\leqslant\frac{\langle w\rangle}{\dfrac{1}{H_c}-\dfrac{1}{H}}, \tag{7}$$

where $H_c=z\{\ln[\rho_c(0)/\rho_c]\}^{-1}$ is the average cloud mass scale height between $z=0$ and $z=z$. We clearly require estimates of the maximum possible values for H_c and $\langle w\rangle$ to obtain the maximum possible values of K.

The discussion thus far has assumed the validity of the Stokes-Cunningham formula [Eq. (3)] throughout the atmosphere. In fact, (3) and therefore (6) and (7) are valid only when the gas mean free path $l<r$. For $r\approx1$ μm this condition is met on Venus for pressures >50 mb. We will be utilizing data pertaining to the 5 to 50 mb region as well as to the region below 50 mb. A formula is therefore required to describe the transition between the "laminar-bulk-flow" drag force used in the Stokes-Cunningham formula and the "molecular-flow" drag force which would pertain to pressures much less than 5 mb. Although no exact theory exists for this transition, an empirical formula for the drag force has been derived experimentally by Millikan (see, e.g., Kennard, 1938). This leads to the following expression for the mass-weighted mean sedimentation velocity for cloud material when $l>r$:

$$\left.\begin{aligned}
\langle w'\rangle&=\frac{\left[\displaystyle\int_0^\infty(\tfrac{4}{3}\pi r^3D)\left(\frac{2gD}{9\rho v(r)}r\right)f(r)dr\right]}{\left[\displaystyle\int_0^\infty(\tfrac{4}{3}\pi r^3D)f(r)dr\right]}\\
&\approx\frac{2gD}{9\rho v(r_m)}r_m\\
v(r)&=\frac{1}{2}\left(\frac{8kT}{\pi m}\right)^{\frac{1}{2}}\Big/\left[\frac{r}{l}+1.23+0.41\exp\left(\frac{-0.88r}{l}\right)\right]
\end{aligned}\right\}. \tag{8}$$

Here k is Boltzmann's constant, m the molecular weight of CO_2, and the "effective sedimentation radius" r_m is identical to the mass-weighted mean radius defined in the next section. For $r_m=1.25$ μm (its maximum possible value), we obtain a maximum possible value for $\langle w'\rangle$ at 50 mb of 0.52 mm sec^{-1}. This can be compared to the maximum possible value of $\langle w \rangle$ at 50 mb (0.43 mm sec^{-1}). In view of the empirical nature of (8) a simple transition from the roughly constant sedimentation velocity defined by (3) below 50 mb to the exponentially increasing sedimentation velocity defined by (8) above 50 mb seems adequate for our purposes.

For $l>r$, $v(r)$ does not vary much with altitude and we can therefore approximate $\langle w' \rangle$ by $\langle w' \rangle(0)\rho(0)/\rho$ where $\langle w' \rangle(0)$ and $\rho(0)$ are the values of $\langle w' \rangle$ and ρ at $z=0$. An analytical solution to (1) and (2) is not possible using this expression for $\langle w' \rangle$. However, if we refer to (6) it is easy to show on general grounds that the minimum possible value for $\rho_c/\rho_c(0)$ is obtained when $t_0=\infty$, both when the sedimentation velocity is roughly constant as in (3) or variable as in (8). We can therefore solve the simpler flux equation

$$K\rho\frac{\partial}{\partial z}\left(\frac{\rho_c}{\rho}\right)+\langle w' \rangle(0)\rho(0)\left(\frac{\rho_c}{\rho}\right)=0, \tag{9}$$

with the boundary condition $\rho_c=\rho_c(0)$ at $z=0$. This then gives the required minimum value for $\rho_c/\rho_c(0)$, namely

$$\rho_c \geqslant \rho_c(0)\exp\left(-\frac{H\Delta\langle w' \rangle}{K}-\frac{z}{H}\right), \tag{10}$$

where $\Delta\langle w' \rangle=\langle w' \rangle-\langle w' \rangle(0)$. In a dust cloud (i.e., $t_0=\infty$) the *local* dust mass scale height from (10) is $[(\langle w' \rangle/K)+(1/H)]^{-1}$. About a scale height above the level where $\langle w' \rangle=K/H$, this local scale height begins to decrease exponentially and ρ_c very rapidly approaches zero. This "dust bank" effect was first pointed out by Hunten in 1970 in an unpublished paper on the production of sodium atoms from micrometeorite dust in the terrestrial atmosphere.

To obtain a comparable expression to (7) for the maximum K value, we combine (10) with the definition of the *average* cloud mass scale height H_c given earlier to yield

$$K\leqslant\frac{H\Delta\langle w' \rangle/z}{\dfrac{1}{H_c}-\dfrac{1}{H}}. \tag{11}$$

The quantity $H\Delta\langle w' \rangle/z$ in (11) defines the "effective sedimentation velocity" for use in the transition region above 50 mb and this can be compared to the effective sedimentation velocity $\langle w \rangle$ used in (7) for the Stokes-Cunningham region below 50 mb. In the limit of very small z, $H\Delta\langle w' \rangle/z=\langle w' \rangle$ as expected. In using (11) for the first few scale heights above the 50-mb level, we see the decrease in H/z partially cancels the nearly exponential increase in $\Delta\langle w' \rangle$ with z. Thus, the difference between the effective sedimentation velocity below 50 mb and that in the 50 to 5 mb region is significantly less than the exponential behavior of $\langle w' \rangle$ might suggest. In the subsequent discussion we will use (3) and (7) below 50 mb and (8) and (11) above 50 mb.

3. Cloud mass scale height

The mass-weighted mean particle radius

$$r_m=\int_0^\infty r^4 f(r)dr \Big/ \int_0^\infty r^3 f(r)dr,$$
$$=\langle r \rangle(1+\delta), \tag{12}$$

is 1.12 μm, which is essentially equal to the extinction radius $\langle r \rangle$. Thus, estimates of the volume particle extinction scale height from optical measurements provide an excellent estimate of the cloud mass scale height H_c. This fortunate result is again due to the very narrow particle size range on Venus. We can conveniently obtain the volume particle extinction scale height using estimates (at a particular wavelength) of two particle extinction optical depths: τ_i at altitude z_i and τ_j at altitude z_j. We therefore write

$$\frac{\tau_i}{\tau_j}=\frac{\sigma[\rho_c(z_i)/(\frac{4}{3}\pi r_m^3 D)]H_c}{\sigma[\rho_c(z_j)/(\frac{4}{3}\pi r_m^3 D)]H_c}=\exp\left(\frac{z_j-z_i}{H_c}\right). \tag{13}$$

Hansen and Arking (1971) and Hansen and Hovenier (1974) have concluded that the mean particle extinction cross section σ varies remarkably little for wavelengths from 0.34 up to 0.99 μm (and perhaps as long as 3.6 μm). We may therefore conceivably use estimates of τ_i and τ_j obtained at two different wavelengths in (13). At present we will assume that r_m is independent of altitude and comment further on this assumption in the next section.

Particle extinction optical depths can be obtained at three different levels in the visible atmosphere on Venus using three distinct observations. Throughout this analysis we will utilize a standard model of the Venus atmosphere (NASA, 1972) based on spacecraft data, in order to relate the temperature, pressure and altitude scales.

a. *Refraction data*

Observations of transits of Venus across the sun have been analyzed by Goody (1967). Using his results and our model atmosphere, we can estimate the vertical cloud extinction optical depth τ_1 at a pressure of $4.9^{+5.1}_{-2.6}$ mb ($z_1=80\pm3$ km). This particular level is where the slant cloud extinction optical depth for visible light (wavelength $\sim$0.5 μm) is approximately unity.[2] We obtain

$$\tau_1=\left(\frac{H_c}{2\pi R}\right)^{\frac{1}{2}}, \tag{14}$$

[2] Ultraviolet photographs of Venus recently obtained by Mariner 10 (Murray *et al.*, 1974) provide excellent confirmation of Goody's analysis.

where $R=6050+z_1$ [km] is the distance from the center of Venus.

b. Polarization data

Hansen and Arking (1971) and Hansen and Hovenier (1974) have concluded that the vertical particle extinction optical depth $\tau_2=1$ at the 50 ± 25 mb level ($z_2=69.5^{+3.5}_{-2}$ km). This result is obtained from analysis of the gas contribution to polarization of reflected solar radiation particularly at wavelengths in the 0.34 to 0.4 μm region. The error estimate on the pressure level given here is due primarily to an estimate by the authors of the effect of proceeding from their homogeneous model to an inhomogeneous one. A pressure of 50 ± 10 mb (69.5 ± 1 km) is suggested from their homogeneous model but we will assume the larger error estimate here to avoid prejudice.

c. Absorption-line data

Several analyses for the 1.05-μm CO_2 band have been carried out using a single isotropically-scattering homogeneous haze layer model (Belton *et al.*, 1968; Belton, 1968; Chamberlain and Smith, 1970). These analyses imply that the product of pressure and specific abundance is $\sim4\times10^3$ (cm amagat) atm.[3] The significance of this particular product is evident from Chamberlain's (1970) theoretical analyses. The cloud-scattering mean free path, λ_3, at the line-forming level (pressure P_3, temperature T_3) can be obtained from this product using

$$\lambda_3=\frac{4\times10^3T_3(1-g)}{273P_3{}^2}. \tag{15}$$

Here we have allowed for anisotropic scattering using the anisotropy coefficient g and the Van de Hulst and Grossman (1968) similarity relations (*cf.* Hansen, 1969; Potter, 1969). At wavelengths near 1.0 μm, estimates of g vary from 0.715 (Hansen and Hovenier, 1974) to 0.85 (Hansen, 1969). The scattering models imply $240\leqslant T_3\leqslant295$K, and $50\leqslant P_3\leqslant200$ mb. Using our model atmosphere to eliminate incompatible P_3, T_3 pairs, we conclude $240\leqslant T_3\leqslant253$K, $110\leqslant P_3\leqslant200$ mb, and $62\leqslant z_3\leqslant65.5$ km. We obtain $0.14\leqslant\lambda_3\leqslant0.264$ km at $z_3=62$ km and $0.44\leqslant\lambda_3\leqslant0.83$ km at $z_3=65.5$ km. Single layer *inhomogeneous* haze models (Regas *et al.*, 1972, 1973) imply $0.11\leqslant\lambda_3\leqslant0.266$ km at $z_3=62$ km ($P_3=200$ mb) in good agreement. *Two-layer* inhomogeneous models have also been shown to be compatible with absorption-line data. These models require very dense clouds at the 200-mb level. For example, Hunt (1972) has $\lambda_3=0.01$ km while Carleton and Traub (1972) and Chamberlain and Smith (1972) use $\lambda_3=0$ km. We therefore have $0\leqslant\lambda_3\leqslant0.83$ km at $z_3=65.5$ km, $0\leqslant\lambda_3\leqslant0.266$ km at $z_3=62$ km, and $0\leqslant\lambda_3\leqslant0.43$ km at $z_3=63.5$ km (155-mb level). We should add parenthetically that two-layer models in themselves require weak vertical mixing to maintain their separate identities.[4]

We can now obtain the particle extinction optical depth at z_3 using

$$\tau_3=H_c/\lambda_3. \tag{16}$$

From (13) and (14) we have

$$\frac{\tau_2}{\tau_1}=\exp\left(\frac{z_1-z_2}{H_c}\right)=\left(\frac{H_c}{2\pi R}\right)^{-\frac{1}{2}}, \tag{17}$$

while from (13), (15) and (16) we have

$$\frac{\tau_3}{\tau_2}=\exp\left(\frac{z_2-z_3}{H_c}\right)=\frac{H_c}{\lambda_3}. \tag{18}$$

Unique solutions may be obtained for the unknowns H_c, τ_1 and τ_3 in (17) and (18). In Table 1 we have listed these solutions taking into full account the quoted error limits on the various quantities. In the 5–50 mb region we find $0.504\leqslant H_c\leqslant2.15$ km and $0.00036\leqslant\tau_1\leqslant0.00075$. In the 50–155 mb region we find $0\leqslant H_c\leqslant4.47$ km and $2.55\leqslant\tau_3\leqslant\infty$. An independent τ_3 estimate can be made using Chamberlain's (1965) formula after scaling for anisotropic scattering (Van de Hulst and Grossman, 1968). For cosines of the incident and emergent angles of radiation to the planetary normal equal to $1/\sqrt{3}$, we obtain

$$\tau_3\approx\{[3(1-\tilde{\omega})]^{\frac{1}{2}}(1-g)\}^{-1}. \tag{19}$$

We caution even when both g and the isotropic single-scattering albedo $\tilde{\omega}$ are known accurately, Eq. (19) provides only a rough estimate of τ_3 (see Chamberlain and McElroy, 1966). For $\tilde{\omega}\approx0.95$ within the moderately strong 1.05-μm CO_2 band, and $g\approx0.8$, we obtain $\tau_3=13$ in rough agreement with our more exact estimates.

We also quote in Table 1 the maximum value for K (K_{max}) defined by (7) or (11). The table suggests very firm upper limits for K of 6.8×10^4 cm^2 sec^{-1} in the 5–50 mb region and 1.1×10^5 cm^2 sec^{-1} in the 50–155 mb region.

For $K=6.8\times10^4$ cm^2 sec^{-1} and $H=5.01$ km, $\langle w'\rangle=K/H$ at 76 km. About a scale height above this level (i.e., $\sim$81 km) we expect the density of a dust cloud on Venus to rapidly approach zero from Hunten's "dust-bank" effect discussed earlier. This appears to be the case on Venus although detailed observa-

[3] Amagat: gas density in units of Loschmidt's number (2.68×10^{19} molecules cm^{-3}).

[4] The Mariner 10 photographs (Murray *et al.*, 1974) in fact confirm the existence of thin highly stratified hazes above the main cloud deck. The considerable atmospheric stability implied by these hazes is in good accord with the results presented here.

TABLE 1. A compilation of values of the maximum allowable vertical mass diffusion coefficient K_{max} ($cm^2 sec^{-1}$) obtained using (7) or (11), the particulate and atmospheric mass scale heights H_c and H (km), and the effective particle sedimentation velocities $\langle w \rangle$ and $H[\langle w' \rangle(z_1) - \langle w' \rangle(z_2)]/(z_1 - z_2) = H\Delta\langle w' \rangle/\Delta z$ (mm sec^{-1}) for various limiting and average values of altitudes z_i (km), pressures P_i (mb), temperatures T_i (°K), extinction optical depths τ_i, and scattering mean free paths $\lambda_i = H_c/\tau_i$. The table implies very firm upper limits for K of 6.8×10^4 $cm^2 sec^{-1}$ in the 5–50 mb region 1.1×10^5 $cm^2 sec^{-1}$ in the 50–155 mb region.

a. The 5–50 mb region

z_1	P_1	T_1	τ_1	λ_1	z_2	P_2	T_2	τ_2	λ_2	H[a]	H_c	$H\frac{\Delta\langle w' \rangle}{\Delta z}$[b]	K_{max}
83	2.3	182	0.00075	2880	67.5	75	233	1	2.15	5.01	2.15	1.81	6.8×10^4
83	2.3	182	0.00059	2265	73	25	211	1	1.34	4.68	1.34	2.45	4.6×10^4
77	10	198	0.00058	2210	67.5	75	233	1	1.27	5.27	1.27	0.72	1.2×10^4
77	10	198	0.00036	1400	73	25	211	1	0.504	4.90	0.504	1.06	6.0×10^3
80[d]	4.9	188	0.00061	2345	69.5	50	227	1	1.42	5.01	1.42	1.30	2.6×10^4

b. The 50–155 mb region

z_2	P_2	T_2	τ_2	λ_2	z_3	P_3	T_3	τ_3	λ_3	H[a]	H_c	$\langle w \rangle$[c]	K_{max}
73	25	211	1	4.47	65.5	110	240	5.32	0.84	5.39	4.47	0.424	1.1×10^5
73	25	211	1	4.03	62	200	253	15.2	0.266	5.47	4.03	0.413	6.4×10^4
67.5	75	233	1	2.14	65.5	110	240	2.55	0.84	5.70	2.14	0.406	1.4×10^4
67.5	75	233	1	2.46	62	200	253	9.27	0.266	5.98	2.46	0.396	1.7×10^4
—	—	—	1	0	—	—	—	∞	0	—	0	—	0
69.5[d]	50	227	1	3.05	63.5	155	248	7.11	0.43	5.75	3.05	0.403	2.6×10^4

[a] From NASA (1972).
[b] Values of $\langle w' \rangle$ computed using Eq. (8) with $r_m = 1.25$ μm, $D = 1.5$ gm cm^{-3}, and l, ρ and T values from NASA (1972).
[c] Maximum values computed using Eq. (3) with $\bar{r} = 1.3$ μm, $D = 1.5$ gm cm^{-3}, and appropriate height-averaged η values from NASA (1972).
[d] Average (preferred) values are given in this row.

tions of particulate distribution above the 76-km level would be required to verify it.

4. Vertical variation of $f(r)$

There is one possible criticism of the procedures we have just followed for the region *below* 50 mb. It might be argued that condensation or particle coagulation could enable a different $f(r)$ at 155 mb than that observed at and above 50 mb. However, we can readily show that even small changes in $f(r)$ with altitude are incompatible with large K values.

The following maximum values are observed at 50 mb: $\langle r \rangle = 1.15$ μm, $r_m = 1.25$ μm, $\bar{r} = 1.3$ μm, $\langle w \rangle \approx 0.43$ mm sec^{-1}, and w_m (the sedimentation velocity of particles with radius r_m) ≈ 0.39 mm sec^{-1}. Let us consider the possibility that $f(r)$ is still given by (5) at 155 mb but the maximum values of the above quantities have increased to $N\langle r \rangle$, Nr_m, $N\bar{r}$, $N^2\langle w \rangle$ and $N^2 w_m$, respectively. In this case the average sedimentation velocity would be $\sim\langle w \rangle(N^2+1)/2$ and the maximum K value implied from (7) would exceed $0.55\times 10^5(N^2+1)$ $cm^2 sec^{-1}$. However, an entirely independent upper limit to K can be derived for this case. Cloud mass density will remain constant or decrease with altitude in the topside of the cloud so we expect $\geqslant N^{-3}$ times more particles with radii Nr_m at 155 mb than particles with radii r_m at 50 mb. The Nr_m radius particles at the 155 mb level ($z_3 = 63.5$ km) will be turbulently mixed up to the 50 mb level ($z_2 = 69.5$ km) where the ratio of the number densities of Nr_m and r_m radius particles must not exceed the observed value defined by (5). We therefore require

$$N^{-3}\exp\left[-(z_2-z_1)\left(\frac{1}{H}+\frac{N^2 w_m}{K}\right)\right] \leqslant N^{(1/\delta)-3}\exp\left[-\left(\frac{Nr_m - r_m}{\delta\langle r \rangle}\right)\right], \quad (20)$$

where $H = 5.75$ km in the 50–155 mb region. For r_m, δ and $\langle r \rangle$ equal to their maximum values and for $1.43 \leqslant N \leqslant 103$ we conclude $K \leqslant 2\times10^5$ $cm^2 sec^{-1}$. For $N < 1.43$, the maximum K value implied from (20) approaches infinity. However, for $N \leqslant 1.43$ the maximum K value implied from (7) is $\leqslant 0.55\times10^5(N^2+1) = 1.7\times10^5$ $cm^2 sec^{-1}$. As N is simply the ratio of r_m (or $\bar{r}$) at 155 mb to r_m (or $\bar{r}$) at 50 mb, Eq. (20) can also be applied to a size distribution at 155 mb typical of coagulating particles [$f(r) = r^{-\beta}$, where $3 \lesssim \beta \lesssim 4$; Junge (1963)], or indeed to any arbitrary $f(r)$ at 155 mb. We can therefore state a firm upper limit for K in the 50–155 mb region of 2×10^5 $cm^2 sec^{-1}$ irrespective of changes in the particle size distribution in this region.

The possibility of an increase in the mean particle size well *above* 50 mb has no effect on our arguments. Such a possibility exists, for example, if HCl and H_2O condense well above 50 mb (as droplets or crystals depending on the temperature). We have argued here that the particles at high altitudes have been turbulently mixed up from the 50-mb level. If an *in situ* particle source exists at high altitudes we will clearly require *smaller* K values in order to satisfy the refraction and polarization data.

5. Discussion and conclusions

We conclude that K on Venus cannot exceed 7×10^4 cm² sec⁻¹ in the 5–50 mb region and 2×10^5 cm² sec⁻¹ in the 50–155 mb region. This estimate is independent of the cloud particle composition or mode of formation, and of changes in the particle size distribution with altitude. Larger values must be considered as completely at odds with current interpretations of refraction, polarization and absorption-line data. The vertical transport time $H^2/K \gtrsim 3.7\times10^6$ sec (5–50 mb), or 1.7×10^6 sec (50–155 mb). These times considerably exceed the horizontal transport time quoted earlier and ensure the validity of our procedure of averaging out the horizontal motions.

The Venera 8 *in situ* wind measurements (Marov *et al.*, 1973) suggest zonal wind velocities $\sim$100 m sec^{-1} at the 1-bar level. Similar zonal wind speeds have been inferred from the Doppler shifts of CO_2 lines formed at the $\sim$155 mb level (Traub and Carleton, 1971), and from the UV photographs which refer to pressures of a few millibars (Rea, 1972). Vertical shear in the zonal wind u may conceivably be very small in the visible atmosphere. We can certainly quote an upper limit for the mean shear in the 5–155 mb region by assuming the zonal wind increases by 100 m sec^{-1} between z_1 and z_3. Thus,

$$\partial u/\partial z \lesssim 6\times10^{-3}\ \text{sec}^{-1}. \tag{21}$$

In addition, the atmosphere becomes abruptly subadiabatic above the $\sim$155 mb level (Fjeldbo *et al.*, 1971). In our NASA (1972) model we estimate the gradient in potential temperature θ in the 5–155 mb region to be expressed by

$$\frac{1}{\theta}\frac{\partial\theta}{\partial z} \approx 3\times10^{-7}\ \text{cm}^{-1}. \tag{22}$$

We can therefore quote a lower limit for the mean Richardson number Ri in this region, namely

$$\text{Ri} = \frac{\dfrac{g}{\theta}\left(\dfrac{\partial\theta}{\partial z}\right)}{\left(\dfrac{\partial u}{\partial z}\right)^2} \gtrsim 7. \tag{23}$$

If we utilize Deardorff's (1967) empirical parameterization for the vertical small-scale eddy diffusion coefficient, K_{eddy}, in the earth's atmosphere, we obtain on Venus

$$K_{\text{eddy}} \lesssim \frac{6\times10^5}{1+50\ \text{Ri}} = 1.7\times10^3\ \text{cm}^2\ \text{sec}^{-1}. \tag{24}$$

Evidently, vertical transport by small-scale eddies is *on the average*[5] quite small in the visible atmosphere.

If vertical transport is by the planetary-scale dynamics the upper limits to K quoted above imply that the vertical wind scale $W = K/H < 1.35$ mm sec^{-1} in the 5–50 mb region and 3.5 mm sec^{-1} in the 50–155 mb region. From continuity, the horizontal wind scale, $U < LW/H = 5.2$ m sec^{-1} in the 5–50 mb region and 11.6 m sec^{-1} in the 50–155 mb region. Here L is taken as the distance from the sub-solar to anti-solar point. The planetary-scale dynamics are certainly not producing a rapid overturn of the visible atmosphere. It is quite conceivable that the visible atmosphere, particularly in the 5–50 mb region, is in a state closely resembling solid (shearless) rotation.

Acknowledgments. I thank Dr. Donald Hunten for his useful comments on an earlier version of this paper and particularly for his advice on the breakdown of the Stokes-Cunningham formula at low pressures. This research was supported in part by the National Aeronautics and Space Administration under Contract NAS2-7994 to MIT.

REFERENCES

Belton, M. J. S., 1968: Theory of the curve of growth and phase effects in a cloudy atmosphere: Applications to Venus. *J. Atmos. Sci.*, **25**, 596–609.

——, D. Hunten and R. Goody, 1968: Quantitative spectroscopy of Venus in the region 8000–11000 Å. *The Atmospheres of Venus and Mars*, J. Brandt and M. McElroy, Eds., New York, Gordon and Breach, 69–97.

Carleton, N. P., and W. A. Traub, 1972: Prediction of atmospheric parameters on Venus from high resolution spectroscopy of a CO_2 line. *Bull. Amer. Astron. Soc.*, **4**, 362 (Abstract).

Chamberlain, J. W., 1965: The atmosphere of Venus near her cloud tops. *Astrophys. J.*, **141**, 1184–1205.

——, 1970: Behaviour of absorption lines in a hazy planetary atmosphere. *Astrophys. J.*, **159**, 137–158.

——, and M. McElroy, 1966: Diffuse reflection by an inhomogeneous planetary atmosphere. *Astrophys. J.*, **144**, 1148–1159.

——, and G. R. Smith, 1970: Interpretation of the Venus CO_2 absorption bands. *Astrophys. J.*, **160**, 755–765.

——, and ——, 1972: Spectrum of Venus with a double-cloud model. *Astrophys. J.*, **173**, 469–475.

Deardorff, J. W., 1967: Empirical dependence of the eddy coefficient for heat upon stability above the lowest 50 m. *J. Appl. Meteor.*, **6**, 631–643.

Fjeldbo, G., A. J. Kliore and V. R. Eshleman, 1971: The neutral atmosphere of Venus as studied with the Mariner 5 radio occulation experiments. *Astron. J.*, **76**, 123–140.

Goody, R., 1967: The scale height of the Venus haze layer. *Planet. Space Sci.*, **15**, 1817–1819.

Hansen, J. E., 1969: Absorption line formation in a scattering planetary atmosphere: A test of Van de Hulst's similarity relations. *Astrophys. J.*, **158**, 337–349.

——, and A. Arking, 1971: Clouds of Venus: Evidence for their nature. *Science*, **171**, 669–672.

[5] The Mariner 10 photographs (Murray *et al.*, 1974) do indicate possible convection in a small region around the sub-solar point. Our results do not rule out such vigorous convection providing it is occurring only over a very small fraction of the planet.

——, and J. W. Hovenier, 1974: Interpretation of the polarization of Venus. *J. Atmos. Sci.*, **31**, 1137–1160.

Hunt, G. E., 1972: Formation of spectral lines in a planetary atmosphere. II. Spectroscopic evidence for the structure of the visible Venus clouds. *J. Quant. Spectros. Radiat. Transfer*, **12**, 405–419.

Junge, C. E., 1963: *Air Chemistry and Radioactivity*. New York, Academic Press, p. 116.

Kennard, E. H., 1938: *Kinetic Theory of Gases*. New York, McGraw-Hill, Section 175.

Marov, M. Ya., V. S. Avduevsky, V. V. Kerzhanovich, M. K. Rozhdestvensky, N. F. Borodin and O. L. Ryabov, 1973: Venera 8: Measurements of temperature, pressure, and wind velocity on the illuminated side of Venus. *J. Atmos. Sci.*, **30**, 1210–1214.

McElroy, M. B., N. D. Sze and Y. L. Yung, 1973: Photochemistry of the Venus atmosphere. *J. Atmos. Sci.*, **30**, 1437–1447.

——, and T. Donahue, 1972: Stability of the Martian atmosphere. *Science*, **177**, 986–988.

Murray, B., M. Belton, G. Danielson, M. Davies, D. Gault, B. Hapke, B. O'Leary, R. Strom, V. Suomi and N. Trask, 1974: Venus: Atmospheric motion and structure from Mariner 10 pictures. *Science*, **183**, 1307–1315.

NASA, 1972: Space vehicle design criteria (environmental) models of the Venus atmosphere. NASA Sp 8001, Washington, D. C.

Parkinson, T. D., and D. Hunten, 1972: Spectroscopy and aeronomy of O_2 on Mars. *J. Atmos. Sci.*, **29**, 1380–1390.

Potter, J. F., 1969: Effect of cloud scattering on line formation in the atmosphere of Venus. *J. Atmos. Sci.*, **26**, 511–517.

Prinn, R. G., 1973: Venus: Composition and structure of the visible clouds. *Science*, **182**, 1132–1135.

Rea, D. G., 1972: Composition of the upper clouds of Venus. *Rev. Geophys. Space Phys.*, **10**, 369–378.

Regas, J. L., L. P. Giver, R. W. Boese and J. H. Miller, 1972: Theoretical interpretation of the Venus 1.05 micron CO_2 band and the Venus 0.8189 micron H_2O line. *Astrophys. J.*, **173**, 711–725.

——, ——, —— and ——, 1973: An expanded theoretical interpretation of the Venus 1.05 micron CO_2 line and the Venus 0.8226 micron H_2O line. *Astrophys. J.*, **185**, 383–390.

Smith, B. A., 1967: Rotation of Venus, continuing contradictions. *Science*, **158**, 114–116.

Traub, W. A., and N. P. Carleton, 1971: Measurements of the differential Doppler effect of a CO_2 absorption line between various points on Venus. *Bull. Amer. Astron. Soc.*, **3**, 278 (abstract only).

Van De Hulst, H. C., and K. Grossman, 1968: Multiple light scattering in planetary atmospheres. *The Atmospheres of Venus and Mars*, J. Brandt and M. McElroy, Eds., New York, Gordon and Breach, 35–55.