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The Dynamics of the Atmosphere of Venus

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ABSTRACT

The wide range of radiative time scales in the Venus atmosphere, together with observations of temperature structure and winds, indicate that the atmosphere contains two distinct regimes. In the deep atmosphere, at altitudes below 40 km, diurnal effects are negligible, motions are weak, and the lapse rate is near-adiabatic. In the upper atmosphere, at altitudes above 70 km, diurnal effects are important, strong retrograde zonal motions \( \sim 100 \text{ m s}^{-1} \) occur, and the lapse rate is sub-adiabatic. The transition region between these two regimes is complicated by the presence of two layers of small-scale turbulence at altitudes of 45 and 60 km.

Analytical and numerical studies show that the Hadley cell hypothesis for the circulations in the deep atmosphere is consistent with all the observations, provided that the greenhouse effect is strong enough to explain the high surface temperatures. Under these conditions the Hadley cell circulation produces an adiabatic, non-turbulent temperature structure, with equator-to-pole temperature contrasts \( \sim 0.1 \text{ K} \), meridional velocities \( \sim 2 \text{ m s}^{-1} \), zonal velocities \( \sim 1 \text{ m s}^{-1} \), and vertical velocities \( \sim 4 \text{ cm s}^{-1} \). Studies of the motions in the upper atmosphere are more ambiguous. Suggestions for explaining the strong zonal motions include the "moving flame" mechanism, the instability of diurnal convective cells to a mean shear, tidal forcing, momentum transport by internal gravity waves, and momentum transport by a Hadley cell. The "moving flame" mechanism has not yet been analyzed for conditions appropriate to the Venus upper atmosphere. The other mechanisms all require at least one ad hoc assumption in order to produce the desired velocities. The location and properties of the two layers of small-scale turbulence suggest the possibility that the lower one is generated by local shear instability, and the upper one by local convective.

1. Observational evidence

During the past decade observational evidence of motions in the Venus atmosphere has steadily accumulated. The first indirect evidence came from observation in ultraviolet (UV) light showing an apparent retrograde rotation rate of 4 days for UV features (Boyer and Camichel, 1961; Boyer and Guérin, 1966). Since the planet’s rotation is retrograde with a period of 244 days, the UV observations imply atmospheric motions in the direction of rotation with a speed \( \sim 100 \text{ m s}^{-1} \). These ground-based observations were confirmed by Mariner 10 (Murray et al., 1974). The latest analyses of the Mariner 10 UV data (Suomi, 1975) show zonal velocities \( \sim 90 \text{ m s}^{-1} \) in low and mid-latitudes and meridional velocities no more than a few meters per second.

Evidence that these strong zonal motions reflect actual mass motions and not merely phase velocities...
comes from both spectroscopic and probe measurements. Spectroscopic observations of Doppler shifts in both reflected sunlight (Guinot and Feissel, 1968) and in carbon dioxide lines formed in the Venus atmosphere (Traub and Carleton, 1975) show retrograde zonal velocities of the order of 100 m s\(^{-1}\). Three of the Venera probes (4, 7 and 8) descended in areas of the Venus atmosphere far enough from the sub-earth points that the probe's drift could be used to deduce horizontal motions in the atmosphere. These measurements showed retrograde zonal velocities as large as 100 m s\(^{-1}\) (Marov et al., 1973).

However, the measurements also indicate that these strong motions are variable. The component of the horizontal velocity measured by the Venera 4, 7 and 8 probes made angles of 70°, 20° and 25°, respectively, with the retrograde zonal direction (Marov et al., 1973). The corresponding maxima in the velocity profiles, all near a height of 50 km, were \(~50\), 10 and 120 m s\(^{-1}\) (Kerzhanovich et al., 1972; Marov et al., 1973). These measurements cannot be reconciled with a flow uniform in direction and magnitude. Similarly the spectroscopic measurements indicate real variations in the flow velocities (Traub and Carleton, 1975). Evidently many more observations will be necessary before we can define the "general circulation" of the Venus atmosphere. The Venera probes did show that the strong velocities are usually confined to the upper atmosphere, to altitudes near 50 km and higher, and that the motions in the lower atmosphere are much weaker, no more than a few meters per second at most in the lowest scale height.

In contrast to the general circulation, the temperature profile in the Venus atmosphere has been well defined by the Mariner 5 observations (Fjeldbo et al., 1971) and the Venera observations (Avduevousk et al., 1970; Marov et al., 1973). Fig. 1 shows the standard model of the Venus atmosphere adapted by NASA (1972), based on the spacecraft observations. The profile is essentially adiabatic below 50 km, all the way to the ground, and this is a sure sign of dynamical activity throughout the lower atmosphere. In fact, the Venera 8 measurement that about 1% of the incident solar flux does reach the ground (Avduevousk et al., 1973) shows that local thermal drives for motions do exist throughout the lower atmosphere. Since the temperature in radiative equilibrium would be roughly proportional to the one-fourth power of the flux, this drive is substantial. Above 65 km the atmosphere is appreciably subadiabatic.

Measurements of thermal emissions from the upper atmosphere (Murray et al., 1963) and of temperatures in the deep atmosphere (Marov et al., 1973) show that horizontal temperature contrasts are quite small, and this is another sure sign of dynamical activity. In the absence of horizontal dynamical transports the horizontal contrasts would be of the same order as the temperatures themselves, as in radiative equilibrium.

The presence of clouds is also an indication of dynamical activity in the atmosphere—either of large-scale rising motions, or of small-scale turbulence. The Mariner 10 UV observations (Murray et al., 1974) show a wealth of detail in the cloud structures. A full inter-
pretation of this detail requires further knowledge of the cloud properties, and it is unclear whether the apparent motions can be identified directly with true motions. However, the patterns themselves provide useful dynamical information. For example, the patterns near the subsolar point strongly suggest small-scale convection (Murray et al., 1974). Other patterns suggest the presence of wave phenomena.

Fluctuations in radar signals transmitted by the Venus atmosphere from the Mariner spacecraft and fluctuations in Doppler shifts of radio signals from the Venus probes are a source of information about small-scale turbulence (Kerzhanovich et al., 1972; Woo et al., 1974; Woo, 1975). The measurements show that two layers of intense turbulence about 10 km thick occur near altitudes of 45 and 60 km, but that there is no turbulence below 40 km. They also indicate that the "outer scale" of the turbulence in the upper layer is about 5 km (Woo, 1975), i.e., that energy is being put into the turbulence on scales of the order of 5 km.

2. Time scales and atmospheric regimes

One valuable approach to understanding atmospheric dynamics is an examination of the time scales characteristic of the important atmospheric processes. The prime process is radiation, since differential solar heating provides the drive for atmospheric motions. One can specify two time scales characteristic of radiative processes. One is the solar time constant, which is essentially the time required for the atmosphere above a pressure level $P$ to cool off if solar heating is stopped, i.e.,

$$
\tau_{\text{sol}} = \frac{rPH}{(\sigma T^4)}
$$

Here $H$ is the scale height, $T$ the temperature at the level $P$, $\sigma$ the Stefan-Boltzmann constant, and $r$ the ratio of specific heats. The second time scale is the radiative relaxation time $\tau_{\text{rad}}$ which is the time required for a perturbation from radiative equilibrium to decay. This latter time is dependent on the vertical scale of the perturbation, but a priori the only vertical scale available is the scale height. If we adopt this scale, then the two time scales, $\tau_{\text{sol}}$ and $\tau_{\text{rad}}$, are essentially identical as shown by Gierasch et al. (1970), and we are entitled to speak of a single time scale, $\tau_{\text{rad}}$, which characterizes the atmosphere's radiative processes at a pressure level $P$. This time scale has been calculated for a CO$_2$ atmosphere by Goody and Belton (1967). In the Venus atmosphere it ranges from about $10^4$ s near the ground to about $10^6$ s at 80 km altitude.

The characteristic time scale for the dynamical response to radiative heating is the advective time scale, i.e., the time it takes the motions to traverse global distances. If $R$ is the planetary radius and $v$ a typical horizontal velocity, then this time scale is

$$
\tau_{\text{dyn}} = R/v.
$$

The one a priori velocity scale we can form from basic atmospheric parameters is

$$
v = (gH)^{1/4},
$$

where $g$ is the acceleration of gravity. This is the velocity scale that would be characteristic of atmospheric temperature structures corresponding to radiative equilibrium, with typical horizontal temperature contrasts $(\Delta T)_H$ of the same order as the mean temperature $T$. In practice the pressure gradients will be smaller than the gradients in radiative equilibrium, and typical velocities will be reduced from the above scale by the factor $[(\Delta T)_H/T]^{1/4}$ (Gierasch et al., 1970). The velocity

![Fig. 2. Ratios of time scales in the Venus atmosphere as a function of altitude.](image-url)
scale given by (3) is also essentially the phase speed of an external gravity wave, or of a sound wave. With the above choice for \( v \), the a priori dynamical time scale for the Venus atmosphere is typically \( 10^6 \) s.

Finally, there is one external time scale of importance, namely, the length of the Venus day \( \tau_{\text{day}} \). Since the absolute rotation period is 244 days and the period of revolution is 225 days, \( \tau_{\text{day}} \) is \( 1.01 \times 10^7 \) s.

The ratios of the above time scales yield information about the relative importance of different processes. For example, the ratio

\[
\frac{\tau_{\text{rad}}}{\tau_{\text{day}}} = \frac{\delta}{\tau_{\text{day}}}
\]

(4)

tells us whether diurnal effects will be important. If \( \delta \gg 1 \), the nighttime is not long enough for appreciable cooling to occur, and diurnal effects will be negligible. Fig. 2 shows the value of \( \delta \), calculated for the NASA standard atmosphere, from Goody and Belton's results for \( \tau_{\text{rad}} \), assuming that one scale height is the characteristic half-wavelength for deviations from radiative equilibrium. For heights less than 56 km, \( \delta > 1 \), and for heights less than 40 km, \( \delta \gg 1 \). Consequently, diurnal effects will be very small in the deep atmosphere, and latitudinal differential heating will dominate in driving the deep motions. In fact, the Venus probes have not found any appreciable diurnal temperature changes in the deep atmosphere.

In contrast, above 56 km, \( \delta < 1 \), and diurnal differential heating may be important. The diurnal effects could be small even here because of the moderating effect on diurnal changes of thermal emissions from the lower atmosphere. However, Ingersoll and Orton's (1974) analysis of the thermal emissions from the upper atmosphere shows that latitudinal and longitudinal contrasts are of the same order of magnitude although the former are larger. Thus we anticipate that both latitudinal and diurnal drives for the motions are important in the upper atmosphere.

The ratio of the dynamical time scale to the length of the day,

\[
\gamma = \frac{\tau_{\text{dy}}}{\tau_{\text{day}}}
\]

(5)
is essentially an inverse Rossby number (ignoring the factor of 2.4 difference between \( \tau_{\text{day}} \) and the absolute rotation period). Therefore \( \gamma \) measures the importance of Coriolis forces in shaping the motions. If \( \gamma \) is small, Coriolis forces will be negligible, and if \( \gamma \gg \text{O}(1) \), they will be important. Fig. 2 shows \( \gamma \) calculated for the NASA standard atmosphere, and since \( \gamma \ll \text{O}(1) \) everywhere, we don't a priori expect Coriolis forces to play a significant role in the dynamics of the Venus atmosphere.

In the deep atmosphere, the solution of the equation for heat conservation will be independent of \( \tau_{\text{day}} \), since \( \delta \gg 1 \). Similarly the solution of the equations of motion will be independent of \( \tau_{\text{day}} \), since \( \gamma \ll \text{O}(1) \). Therefore the precise value of \( \tau_{\text{day}} \) should be immaterial to the state of the lower atmosphere, and the controlling dimensionless parameter should be that combination of \( \delta \) and \( \gamma \) which is independent of \( \tau_{\text{day}} \), i.e.,

\[
\frac{\tau_{\text{dy}}}{\tau_{\text{sol}}} = \frac{\delta}{\gamma}
\]

(6)

This is the same parameter that Golitsyn (1970) deduced would be the controlling parameter for a non-rotating planet. He postulated that the important external parameters were the amount of solar radiation absorbed per unit area \( (q_A) \), the specific heat of the atmosphere \( (c_p) \), the mass of the atmosphere per unit area \( (M) \), the planetary radius \( (R) \), and the Stefan-Boltzmann constant \( (\sigma) \). From these parameters he formed a single dimensionless parameter:

\[
\frac{q_A^{1/8} R}{c_p^{1/2} \sigma^{1/8} L^{1/8}} = \frac{\sigma^{1/8} R}{c_p^{1/2} L^{1/8}}
\]

(7)

If we note that

\[
M = \frac{P}{g}
\]

(8)

\[
q_A \sim \sigma T^4,
\]

(9)

\[
r - 1 = gH - c_pT = c_pT,
\]

(10)

we can verify by direct substitution that

\[
3M \sim \epsilon.
\]

(11)

Gierasch et al. (1970) showed that \( \epsilon \gg 1 \) implies that the dynamical fluxes of heat are very inefficient compared to the radiative fluxes, and in this case the temperature structure would be that appropriate to radiative equilibrium. If \( \epsilon \ll \text{O}(1) \), then the radiative equilibrium temperature structure will be appreciably modified by the dynamical transports. Fig. 2 shows \( \epsilon \) as a function of height, also calculated for the NASA standard atmosphere. Since \( \epsilon \ll \text{O}(1) \) everywhere, we anticipate that the motions are crucial in determining the temperature structure throughout the atmosphere, and this is of course confirmed by the very small temperature contrasts observed.

The results displayed in Fig. 2 lead us to a very natural separation of the atmosphere into two regimes: a lower atmosphere where \( \delta \gg 1 \), and an upper atmosphere where \( \delta < 1 \). In the NASA standard model, the division occurs at about 56 km. In the lower atmosphere the dominant drive for the motions is differential latitudinal heating, while in the upper atmosphere both differential latitudinal and diurnal heating are impor-
tant. Correspondingly a single parameter \( \epsilon \) controls the dynamical regime in the lower atmosphere, and two \((\epsilon \text{ and } \delta, \text{ or equivalently } \gamma \text{ and } \delta)\) control the dynamical regime in the upper atmosphere. Of course, the transition between these two regimes will not be sharp.

It is interesting to note that this \textit{a priori} separation of the atmosphere shows up in the observations. For example, the changeover from an adiabatic lapse rate to a sub-adiabatic lapse occurs near the level where \( \delta = 1 \) (cf. Figs. 1 and 2), and the strong zonal motions are confined to the regions where \( \delta \leq O(1) \). The two turbulent layers at 45 and 65 km are located in the transition region, and this is likely to be the most difficult region to analyze.

It is also convenient to define the “deep” atmosphere as the levels where \( \delta \gg 1 \). From Fig. 1 we see that this is essentially the portion of the atmosphere below 40 km. The deep atmosphere is the part of the atmosphere that is likely to have a two-dimensional structure, since the diurnal influence is negligible when \( \delta \gg 1 \).

3. Planetary-scale motions in the lower atmosphere

The first theoretical discussion of the circulations in the lower atmosphere was presented by Goody and Robinson (1966). Their basic hypothesis was that, in an atmosphere with negligible rotation subject to differential solar heating, the general circulation would consist of a simple cellular overturning, with rising motions near the subsolar point and sinking motions near the antisolar point. Since we now know that the thermal inertia of the deep atmosphere is so large that \( \delta \gg 1 \), this picture has to be modified so that the rising motions are placed near the equator and the sinking motions near the poles. This kind of motion is generally referred to as a Hadley cell.

Fig. 3 illustrates schematically the streamlines in a Hadley cell. The asymmetric form of the cell, with the region of the rising motions being much broader than the region of sinking motions, is a general feature of Hadley cells, when dynamical transports of heat appreciably affect the temperature structure (Stone, 1968). Since the hypothesis that the circulation is a Hadley cell is essentially a straightforward statement that the motions will be in a thermodynamically direct sense, and since such motions are observed in analogous experimental situations (Rossby, 1965), it is difficult to argue that the general circulation in the lower atmosphere will not be a Hadley cell, at least in an average sense. In fact, no other suggestion has been put forward for the planetary-scale motions in the lower atmosphere, and all discussions of these motions have adopted the Hadley cell picture, implicitly or explicitly.

In their original discussion Goody and Robinson suggested that such a Hadley cell might transport heat downward and account for the high surface temperatures near 750 K. However, this suggestion cannot be reconciled with the thermodynamically direct nature of a Hadley cell. By definition warm air rises and cool air sinks, so that the net heat transfer is upward. In fact, it is generally difficult to explain the high surface temperatures by invoking dynamical transports. Such an explanation requires a thermodynamically indirect circulation, even though the circulations are thermally driven. The most plausible explanation appears to be the greenhouse effect (Sagan, 1962; Pollack, 1969), particularly in view of the Venera 8 flux measurements.

One other qualitative feature of a Hadley cell is important for understanding the lower atmosphere. The cell transports heat meridionally as well as vertically, and the direction of the meridional transport will depend on the static stability of the atmosphere. The cell transports heat poleward only if the poleward branch of the cell at higher levels is on the average at a higher potential temperature than the equatorward branch at lower levels. Thus the Hadley cell hypothesis is consistent with the small temperature contrasts observed between the equator and the pole only if the lapse rate in the deep atmosphere is at least slightly sub-adiabatic, on the average. A super-adiabatic lapse rate would imply an equatorward transport of heat.

One approach used to analyze the lower atmosphere theoretically has been applications of scaling analysis in order to deduce qualitative information about the dynamics and structure of the lower atmosphere without having to solve the differential equations in detail. All these analyses assume that the motion is two-dimensional and therefore they are only applicable to the deep atmosphere. The first two scaling analyses were those presented by Goody and Robinson (1966) and Stone (1968). Both of these assumed that all of the solar radiation was absorbed near the top of the lower atmosphere, and that the large-scale dynamical fluxes at lower levels were balanced by small-scale turbulent fluxes. This led to a picture of a thermodynamically direct Hadley cell with a strong boundary layer next to the layer where the solar radiation was...
absorbed. This picture contains no mechanism for producing the high surface temperature, and therefore is implausible. Nevertheless, these analyses did show that reasonable meridional velocities, \( \sim 30 \text{ m s}^{-1} \), would give rise to meridional heat transports sufficient to account for the very small latitudinal temperature contrasts observed.

A more plausible kind of balance is that assumed in the scaling analyses presented by Gierasch et al. (1970) and by Stone (1974). Here it was assumed that dynamical cooling and heating were balanced by radiative heating and cooling. Gierasch et al. assumed that the typical vertical contrast of potential temperature, \( (\Delta \theta)_v \), was the same magnitude as the horizontal contrast, \( (\Delta \theta)_H \). Then the requirement that the dynamical and radiative heating be in balance globally was used to derive an equation for \( (\Delta \theta)_H \). This analysis demonstrated the dynamical significance of the parameter \( \epsilon \) for the lower atmosphere, and showed that the small values of \( \epsilon \) implied that \( (\Delta \theta)_H \sim 0.2 \text{ K} \), a value consistent with the Venera measurements.

Gierasch et al.'s scaling analysis was extended by Stone (1974), who relaxed the assumption \( (\Delta \theta)_H \sim (\Delta \theta)_v \). By requiring that the dynamical and radiative heating be in balance when averaged either vertically or latitudinally, he obtained two simultaneous equations for \( (\Delta \theta)_H \) and \( (\Delta \theta)_V \). The solution depends on the value of \( (\Delta \theta)_V \) for the radiative equilibrium state, i.e., on the vertical distribution of solar heating. Since this quantity in the lower atmosphere is unknown, he found solutions as a function of the radiative equilibrium value of \( (\Delta \theta)_V \). These solutions for the vertical and latitudinal contrasts in potential temperature are illustrated in Figs. 4 and 5, respectively. They were calculated for \( \epsilon = 10^{-5} \), an appropriate value for the deep atmosphere (see Fig. 1).

Fig. 4 shows that the stabilizing effect of the upward heat transport by the Hadley cell has almost the same effect on the lapse rate as small-scale convection. If the radiative state is statically stable, the lapse rate is essentially the lapse rate corresponding to radiative equilibrium. If the radiative state is statically unstable, the lapse rate becomes essentially adiabatic. In the latter case the lapse rate is actually sub-adiabatic, but by an amount too small to have been measured by the Venera probes. For example, if the lapse rate of the radiative state were \( 12 \text{ K km}^{-1} \), corresponding to \( (\Delta \theta)_V / T \approx -0.1 \), then the Hadley cell would produce a vertical contrast \( (\Delta \theta)_V / T \approx 4 \times 10^{-4} \), corresponding to a static stability \( \sim 0.01 \text{ K km}^{-1} \). Therefore the observed adiabatic lapse rate in the lower atmosphere can be attributed simply to the action of a Hadley cell, without invoking small-scale convection. An explanation of the adiabatic lapse rate based on the large-scale motions rather than small-scale convection is particularly attractive in view of the failure of the Venera probes to detect any small-scale turbulence in the deep atmosphere.

Fig. 5. Equator-to-pole temperature contrast produced by a Hadley cell circulation as a function of the static stability of the radiative equilibrium state.
Fig. 5 shows that a Hadley cell can be relied on to produce extremely small equator-to-pole temperature contrasts in the deep atmosphere. In order of magnitude these contrasts are much smaller when the static stability is large, since the potential temperature difference between the poleward and equatorward branches of the Hadley cell is then much greater, and the motions are much more efficient at transporting heat poleward. Similarly, the stabilizing effect of the vertical flux of heat by the motions is much more efficient when the equator-to-pole contrasts are much larger, since the temperature difference between the warm rising air and the cool sinking air is much greater. From Fig. 5 we see that typical equator-to-pole temperature contrasts in the adiabatic regime are only ~0.1 K. This very small contrast and the adiabatic nature of the lapse rate are both signs of the great efficiency of the dynamical fluxes when $\epsilon$ is as small as $10^{-5}$. The corresponding velocity scales are $2 \text{ m s}^{-1}$ for the meridional velocity and $0.1 \text{ cm s}^{-1}$ for the vertical velocity (Stone, 1974). These scales are all consistent with the Venera measurements in the deep atmosphere.

Since $2 \text{ m s}^{-1}$ is considerably less than the a priori velocity scale given by Eq. (2), the corresponding value of $\gamma$ for the deep atmosphere needs to be re-calculated. The value of $\gamma$ corresponding to $v=2 \text{ m s}^{-1}$ is in fact $1/2$. Since $\gamma$ is still less than unity, Coriolis forces will not change the results of the scaling analysis quoted above. However, $\gamma$ is now large enough that Coriolis forces will produce significant zonal velocities in the deep atmosphere. In particular, we may anticipate zonal velocities $\sim 1 \text{ m s}^{-1}$.

Besides the scaling analyses, there have been a number of investigations of the lower atmosphere circulations which directly integrate the conservation equations of mass, momentum and energy by numerical means. Such investigations have been reported by Hess (1968), Sasmori (1971), Turikov and Chalikov (1971) and Kálnay de Rivas (1973). In principal, these investigations can give us much more detailed information about the lower atmosphere.

Turikov and Chalikov used a two-layer model in which the lower layer was allowed to respond instantaneously to radiative heating, i.e., they took $\delta=0$. As a result they found diurnal circulations to be dominant. In view of the very long thermal time constant of the lower atmosphere, corresponding to $\delta \gg 1$, their results are not very meaningful for the deep Venus atmosphere.

Hess, Sasmori and Kálnay de Rivas all assumed that the flow was two-dimensional, and used a time-marching procedure to integrate the equations. An arbitrary initial state was specified, and the subsequent history of the system calculated. When the system appeared to have evolved to a steady state, the integration was stopped. In all cases this happened after a time $\sim 2 \times 10^5 \text{ s}$. Since this time is short compared to the radiative relaxation time of the deep atmosphere, $10^9 \text{ s}$, Kálnay de Rivas (1973) pointed out that it was questionable whether any of these calculations had in fact reached an equilibrium state. The time development of a Hadley-cell circulation was studied by Stone (1974), and he found that the adjustment time depended on the strength of the radiative heating in the lower atmosphere. Hess, Sasmori, and Kálnay de Rivas all assumed distributions of the solar heating that would be insufficient to produce the high surface temperatures by means of the greenhouse effect, and in this situation Stone (1974) found that the required adjustment time was in fact the radiative relaxation time. Therefore, all these experiments do not simulate equilibrium conditions in the deep atmosphere. For example, all of them found lapse rates close to adiabatic at the end of their integrations, but this is an artifact of their having specified adiabatic lapse rates in their initial conditions.

Recently Kálnay de Rivas (1975) extended her calculations to allow for this long adjustment time, and in fact found that the lower atmosphere gradually cooled off, and the adiabatic lapse rate gradually disappeared. She also described a new, more realistic calculation in which the greenhouse effect was assumed to be strong enough to cause the high surface temperatures. In this calculation the adiabatic lapse rate was maintained; equator-to-pole temperature contrasts $\sim 0.1 \text{ K}$ developed; meridional velocities developed which varied from $\sim 1 \text{ m s}^{-1}$ near the surface to $\sim 5 \text{ m s}^{-1}$ near the top of the lower atmosphere; and vertical velocities developed which varied from $\sim 1 \text{ cm s}^{-1}$ in the rising branch of the Hadley cell to $\sim 5 \text{ cm s}^{-1}$ in the descending branch. Rotational effects were included, and they produced zonal velocities $\sim 10 \text{ m s}^{-1}$, confined to polar regions near the top of the atmosphere. These numerical results are all in good agreement with the predictions of scaling analysis for the deep atmosphere. One important restriction on the numerical results for the upper parts of the lower atmosphere (altitudes $>40 \text{ km}$) is that they were derived with a two-dimensional model. Since $\delta \ll 1$ in these regions, diurnal effects could significantly modify the results in these layers.

4. Planetary-scale motions in the upper atmosphere

The dynamical studies of the upper atmosphere have generally focused on the strong zonal winds. A review of these studies has been presented recently by Young and Schubert (1973). Because of the small values of $\delta$ in the upper atmosphere, most of these studies have examined the effects of diurnal heating, and in fact have used two-dimensional models which exclude latitudinal heating. The Mariner 10 observations (Murray et al., 1974) of an apparent subsolar convection region do indicate that diurnal heating is significant in the upper atmosphere. However, the moderating influence of the lower atmosphere makes it unlikely that latitudinal heating can be neglected. In any case it is valuable to try to understand the simpler kinds of circulation first.
The first mechanism proposed to explain the 4-day retrograde rotation rate, and the one that has been most thoroughly explored, is Schubert and Whitehead's (1969) "moving flame" mechanism. This mechanism in its simplest form explains why a flame moving in a circular path beneath a pan of fluid will induce a motion of the fluid in the opposite direction (Fultz et al., 1959; Davey, 1968). The moving flame induces a thermal wave which lags behind the moving flame by an amount which increases with height above the bottom of the pan, because of the finite time required for conduction to transmit the heat into the fluid. The tilting isotherms will drive tilting convection cells which also lag behind the moving flame at finite heights above the bottom of the pan. Because of this tilt, motions of the fluid directed opposite to the motion of the flame are correlated with rising motions, and there is an upward eddy transport of momentum directed opposite to the flames' motion. Consequently, a net motion in the opposite direction appears at the top of the fluid. Fig. 6 shows schematically the tilted cells and the net zonal motion they will produce in the upper levels, if the atmosphere is heated from below.

The situation in the Venus atmosphere is not so simple. There the thermal lag associated with solar heating is greatest at lower levels, and this would reverse the tilt of the convection cells, and drive a prograde zonal motion in the upper levels. Analyses of this situation by Malkus (1970) and Gierasch (1970) appeared to show retrograde zonal flows, but their solutions are not physically realistic. Malkus' solution contains a mathematical error in sign. Gierasch's solution applies to an inviscid fluid, and is unstable if an infinitesimal amount of viscosity is added. The viscosity would cause a decrease in the shear of the zonal flow, resulting in a tilt to the convection cells which would further decrease the zonal flow because of the eddy transport generated. Other processes have to be invoked to explain a tilt consistent with the observed retrograde motion.

Young and Schubert (1973) have shown that the stratification of the upper atmosphere associated with the sub-adiabatic lapse rate can cause the proper tilt. To solve the equations of motion, they used the mean-field approximation, in which the self-interactions of the convection cells are neglected. The solution is a function of two dimensionless parameters which can be simply related to the ratios of time scales discussed in Section 2. Fig. 7, taken from Fig. 5 of Young and Schubert (1973), shows the magnitude of the mean retrograde zonal motion produced by the tilted convection cells as a function of the parameter $\Lambda$:

$$\Lambda = \frac{1}{15\gamma^2}$$

Young and Schubert found solutions for two values of the radiative relaxation time, corresponding to $\delta = 3.2$ and 0.16, as indicated in Fig. 7. The latter value is a realistic one for the upper atmosphere (see Fig. 2).

Young and Schubert were not able to solve the equations for large values of $\Lambda$ because their iterative technique failed to converge. Unfortunately, the value of $\Lambda$ appropriate to the upper atmosphere of Venus, as indicated in Fig. 7, is an order of magnitude larger than the values for which they were able to obtain solutions.
The trend of their solutions is encouraging and suggests that zonal motions $\sim 100$ m s$^{-1}$ might be attained for the proper value of $\Lambda$. However, extrapolating their solutions is risky. The breakdown of an iterative technique is often a sign of a basic change in the physical mechanisms at work. In addition, even their solutions for the larger values of $\Lambda$ (the dotted parts of the curves in Fig. 7) are not quantitatively accurate, since self-interactions of the convective cells are not negligible for these values of $\Lambda$. It still remains to be demonstrated that the moving flame mechanism can produce $100$ m s$^{-1}$ velocities under conditions appropriate to the upper Venus atmosphere.

Thompson (1970) has suggested another mechanism closely related to the moving flame mechanism. He proposed that tilting convection cells driving the zonal motion could be the result of instability. If initially the convection cells are not tilted, and then a small perturbation in the form of a mean zonal flow with shear is superimposed, the shear will tilt the cells in a direction such as to produce an eddy momentum flux which reinforces the initial perturbation (see Fig. 6). Thompson's numerical calculations of this effect were inconclusive, since his solutions showed instabilities only in cases where his spatial resolution was inadequate to resolve the flow boundary layers. A more careful analysis of the problem by Busse (1972) indicates that this kind of instability will not occur in atmospheres which are shallow compared to the planetary radius. However, Busse's results are also not conclusive for Venus, since they are based on assumptions that may not be relevant, e.g., the assumption of vanishingly small Prandtl number. In any case Thompson's mechanism cannot independently explain the direction of the motion.

Another proposal, put forward by Gold and Soter (1971), is that the solar tidal forces, acting on the second harmonic of the atmospheric mass distribution associated with the semi-diurnal thermal tide, will accelerate the atmosphere to produce the observed zonal motions. The sign and magnitude of the resulting zonal winds depend on the amplitude and phase of the semi-diurnal tide and the effective kinematic viscosity of the atmosphere. None of these quantities are known for Venus, and no theories for predicting them have been proposed. Consequently it is difficult to assess the plausibility of Gold and Soter's mechanism.

Another mechanism has been discussed by Lindzen (1973) and Fels and Lindzen (1973). They pointed out that internal gravity waves generated by diurnal heating will tend to carry prograde momentum out of the layers where solar radiation is absorbed, thereby accelerating these layers in a retrograde direction. The higher and lower layers cannot be accelerated to prograde velocities exceeding the apparent speed of the sun, $4$ m s$^{-1}$, because of critical layer absorption of the waves; and the retrograde acceleration of the radiatively absorbing layers will be counteracted by turbulent dissipation when the Richardson number of the shear layer between the prograde and retrograde layers falls below $\frac{1}{4}$. Fels and Lindzen showed that these constraints, acting under the statically stable conditions of the Venus upper atmosphere, would allow this mechanism to support retrograde velocities $\sim 100$ m s$^{-1}$ only if mean velocities $\sim 25$ m s$^{-1}$ could be produced independently, by some other mechanism. They also found that the resulting region of retrograde motion would be $\sim 5$ km thick. Unfortunately, the observations indicate that the layer with strong zonal velocities is much
thicker than this. The strong velocities appear not only in the UV observations which refer to heights near 70 km, but also in the spectroscopic observations which refer to heights near 60 km, and in the Venera 8 observations which refer to heights near 50 km.

Recently Leovy (1973) and Gierasch (1975) have discussed a mechanism which contrasts strongly with all of the above mechanisms, in that it relies on meridional pressure gradients rather than zonal pressure gradients. In view of the small values of $\delta$ in the upper atmosphere, both kinds of gradients are likely to be present. Leovy suggested that the zonal winds are in "cyclostrophic" balance—i.e., that centrifugal force is balanced by meridional pressure gradients—and that a Hadley cell motion is likely to be associated with such gradients. The small meridional velocities found in the Mariner 10 UV observations (Suomi et al., 1975) imply that the meridonal accelerations are much weaker than the centrifugal forces, so that cyclostrophic balance is plausible.

Since the zonal motions have higher angular momentum in low latitudes than in high latitudes (Suomi et al.), the vertical motions associated with a Hadley cell will transport momentum upward on balance. This transport could support strong zonal velocities in the upper atmosphere. However, at the same time the Hadley cell transports angular momentum poleward. Consequently one needs a very strong horizontal diffusion mechanism to maintain the strong zonal velocities in low latitudes. Gierasch demonstrated that the observations could be matched if the horizontal diffusion of momentum in the upper atmosphere was sufficiently strong that the horizontal kinematic eddy viscosity was much greater than $10^{7} \text{m}^{2}\text{s}^{-1}$. Numerical calculations by Kaidnay de Rivas (1975) show that this mechanism will not work if the eddy diffusivity is of order $10^{5} \text{m}^{2}\text{s}^{-1}$ or larger, since this strong a diffusivity leads to very small temperature gradients and correspondingly weak large-scale motions. Consequently, the mechanism will work only in the rather special case that there is a horizontal mixing mechanism present which is very efficient at diffusing momentum, but orders of magnitude less efficient at diffusing heat.

5. Small-scale instabilities

The presence of adiabatic lapse rates in the lower atmosphere raises the possibility that small-scale convection may occur there, especially in low latitudes where the solar heating is a maximum. However, the Venera probes, which did land in low latitudes, were not able to detect any turbulence below 40 km (Kerzhanovich et al., 1972) and this suggests that convection does not in fact occur in the deep atmosphere. Such a conclusion is supported by the prediction that a Hadley cell will produce a temperature structure which is not only sub-adiabatic in the mean, but also has latitudinal temperature variations which are smaller than the mean static stability (cf. Figs. 4 and 5), so that even local convection would not occur.

There remains the possibility that heating in the upper atmosphere can produce small-scale convection. Calculations of the deposition of solar radiation (Lacis, 1975) show that the strongest heating rates will be near the 60 km level and higher. At the higher levels the lapse rates appear to be sub-adiabatic (see Fig. 1), and therefore convection is most likely to be occurring near the 60 km level. In fact, one of the turbulent layers observed by the Mariner spacecraft was centered near 60 km. This coincidence suggests that this layer may be produced by local small-scale convection. This interpretation is supported by Woo's (1975) analysis, which indicates that the turbulence is isotropic with an outer scale size of 5 km, just as one would expect for small-scale convection in a region with a scale height of 5 km. The convection pattern apparent near the subsolar point in the Mariner 10 UV observations probably refers to higher levels, and may be a sign of the lower level convection penetrating into the upper stable layers above the regions where the diurnal heating is strongest.

In regions of static stability, shear instability can occur if the Richardson number, $\text{Ri}$, is less than one-fourth. Using the Venera 4 measurements, Kerzhanovich et al. (1972) estimated that small values of $\text{Ri}$ do occur in the levels between 40 and 50 km, and suggested that this was the cause of the turbulent layer near 45 km observed by Venera 4. The large uncertainty in the static stability at these levels makes such estimates very uncertain. However, this is just the region where the wind shear measured by Venera 4 was strongest, so shear instability is a plausible mechanism for this turbulent layer. Such an interpretation is also supported by the failure of Venera 7 to find this turbulent layer, when the winds were much weaker.

Prinn (1974) estimated that $\text{Ri} \geq 7$ in the regions of the upper atmosphere where the lapse rates are strongly sub-adiabatic. Consequently, shear instability as well as static instability are unlikely to occur in these higher levels. Prinn also showed that the observed properties of the cloud particles in these levels do not require strong turbulence to support the particles, and this is consistent with the apparent strong dynamical stability of the upper atmosphere.

Hart (1972) has demonstrated that another, weaker kind of instability can occur in statically stable conditions, even if $\text{Ri} > \frac{1}{4}$. This instability depends on the presence of a temperature gradient parallel to the direction of the shear flow. Since Coriolis forces are negligible on Venus, such gradients are likely to exist in the Venus atmosphere. In these circumstances potential energy as well as kinetic energy is available for small-scale disturbances to feed on. In fact, Hart found that instability would occur for any value of $\text{Ri}$.

Hart's calculations showed that the growth rate for
these instabilities, in order of magnitude, is

\[
\frac{\partial u}{\partial T} - \frac{\partial \theta}{\partial \omega} = 0,
\]

where \( \partial u/\partial z \) is the vertical shear of the mean flow, \( \partial T/\partial x \) the horizontal temperature gradient in the direction of the mean flow, and \( \partial \theta/\partial z \) is the static stability.

To estimate a typical growth rate for the upper atmosphere we choose \( \partial u/\partial z \approx 10^{-2} \text{ s}^{-1}, \partial T/\partial x \approx 10^{-4} \text{ K km}^{-1}, \partial \theta/\partial z \approx 5 \text{ K km}^{-1} \). We find \( \omega \approx 2 \times 10^{-6} \text{ s}^{-1} \), corresponding to a time scale \( \approx 5 \text{ days} \). This is rapid enough compared to the other time scales (e.g., the 4-day rotation) to suggest that this instability could play a role in the dynamics of the upper atmosphere. Hart's calculations indicated that the dominant modes when \( R_l > \frac{1}{2} \) would be “finger” modes, i.e., rolls with their axes parallel to the mean flow. Their horizontal wavelengths are comparable to the intrinsic vertical scale, i.e., to the scale height. This instability is one possible explanation for the small-scale zonal streaks apparent in some of the Mariner 10 cloud pictures. They draw energy from both the potential and kinetic energy of the mean state, and thereby tend to destroy the shear of the mean flow and the horizontal temperature gradients, and tend to stabilize the lapse rate.

6. Summary

The wide range of values of \( \delta \) in the Venus atmosphere, together with observations of temperature structure and motions, indicate that the atmosphere contains two distinct regimes. In the deep atmosphere, below 40 km, diurnal effects are negligible, the lapse rate is near-adiabatic, and the motions are weak. In the upper atmosphere, above 70 km, diurnal effects are important, the lapse rate is sub-adiabatic, and zonal motions are strong.

Theoretical attempts to understand the dynamics of the deep atmosphere are in good agreement with the observational data. The Hadley cell hypothesis is capable of explaining the adiabatic, nonturbulent structure, the virtual lack of horizontal gradients, and the observed magnitude of the horizontal velocities. The crucial assumption of the Hadley cell hypothesis is that the greenhouse effect is sufficient to cause the large surface temperatures. In view of the Venera 8 flux measurements, this is not an unreasonable assumption, but we need measurements of net short- and long-wave fluxes to answer the question conclusively. Also it would be valuable to have sufficient measurements to determine the mean circulation in the deep atmosphere directly. This is likely to be more informative than measurements of the mean temperature structure, since the dynamically significant temperature differences appear to be extremely small (cf. Figs. 4 and 5).

The theoretical attempts to understand the dynamics of the upper atmosphere are more ambiguous. None of the mechanisms proposed for explaining the 4-day retrograde motion is completely satisfactory. Young and Schubert's calculations for the “moving flame” mechanism need to be extended to values of \( \lambda \) appropriate to Venus (cf. Fig. 7). All the other proposed mechanisms require at least one ad hoc assumption in order to produce the desired velocities. Also, all discussions of the dynamics of the upper atmosphere have assumed that the motions are two-dimensional, and this is likely to be an oversimplification. Several of the proposed mechanisms may be at work simultaneously. The apparent variability of the observed motions indicates that observations over a long period of time will be necessary to define the general circulation of the upper atmosphere.

The transition region between the deep atmosphere and the upper atmosphere has not yet been the subject of any theoretical investigations. Its structure is quite complicated, as evidenced by the presence of two distinct layers of turbulence at 45 and 60 km. The location and properties of these layers suggest the possibility that the upper layer is generated by small-scale convection and the lower layer by local shear instability.

As the complexity of our observational knowledge of the Venus atmosphere increases, numerical general circulation models will become increasingly important in analyzing the large-scale dynamics. They may be the only feasible way of investigating the three-dimensional, nonsteady motions that occur in the upper atmosphere, and the motions in the transition region with its complicated, partly turbulent structure. Two-dimensional models are of value for studying the deep atmosphere, and numerical calculations for the deep atmosphere should be updated as more information about radiative heating in the lower atmosphere becomes available.

The wealth of detail in the Mariner UV observations invites theoretical investigation of the small-scale motions in the upper atmosphere. “Finger” instability is a possible mechanism for producing the observed small-scale zonal cloud streaks. The wave-like patterns apparent in the observations suggest that it would be particularly worthwhile to investigate the propagation characteristics of the upper atmosphere.

REFERENCES


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