VENUS II

GEOLOGY, GEOPHYSICS,
ATMOSPHERE, AND SOLAR WIND
ENVIRONMENT

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Editors

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THE UNIVERSITY OF ARIZONA PRESS
TUCSON
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VENUS II

GEOLOGY, GEOPHYSICS,

ATMOSPHERE, AND SOLAR WIND

ENVIRONMENT
THE GENERAL CIRCULATION OF THE VENUS ATMOSPHERE: AN ASSESSMENT

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The overall spin or "superrotation" of the Venus atmosphere is a striking phenomenon. In the 15 years since the NASA Pioneer Venus mission, a first-order understanding has been reached of the dynamics of the atmospheric region near and just above the Venus cloud tops. Tidal motions induced by solar heating produce a traveling disturbance whose vertical momentum transports are balanced by mean flow advection. The balance explains the shear of the mean flow above the clouds, and partially explains the strength of the mean flow at the cloud level where the strongest superrotation of
the atmosphere occurs. But the fundamental cause of the global superrotation remains a mystery in spite of data from Earth-based observatories, from Pioneer Venus, from several Russian probes, from a Russian/French balloon experiment, and from the NASA Galileo flyby. The key missing knowledge is of momentum transfer processes in the deep atmosphere, between the surface and the cloud deck. Neither the forcing nor the drag and dissipation mechanisms are known. The existing data are reviewed here, and theoretical suggestions are listed. It is concluded that further measurements, in conjunction with numerical modeling, will be required to resolve this puzzling and challenging question. New data must improve by an order of magnitude on the accuracies achieved by the Pioneer Venus probes. Velocities in the deep atmosphere must be measured to better than 0.1 m s⁻¹, and relative temperatures to better than 0.1 K near the surface.

I. INTRODUCTION

The rotation of the Venus atmosphere is one of the most intriguing unexplained phenomena in planetary science. At cloud-top level, where the pressure is about 50 mb, the mean rotation period is about 4 terrestrial days, in contrast to the 243 day rotational period of the solid planet. The atmosphere is primarily CO₂, and the surface pressure is about 90 bar. The total atmospheric mass is roughly equivalent to 1 km of water, and as a consequence the thermal time constant (evaluated in Sec. III) is very long, approximately one century. Through most of its depth the atmosphere is stably stratified and it is likely that frictional dissipation is weak. In comparison, the relevant flow time constants based on advection are a few days. Thus the Venus atmosphere is weakly damped and weakly forced, and the mean flow may be only loosely coupled to the fundamental drives. The rapid mean flow may be the end state of kinetic energy transformations that are several steps removed from the solar heating that drives the system. It is an interesting system on general dynamical grounds, as well as presenting a puzzle specifically to the atmospheric and planetary sciences.

Our purpose is an assessment of the understanding of the Venus circulation rather than a detailed review of work that has been done, although we attempt to be complete in describing the observational basis. A pre-Pioneer Venus assessment was presented by Hunten and Goody (1969). Detailed post-Pioneer Venus reviews are by Moroz (1981), covering general Venus atmospheric science, and Schubert (1983) covering dynamics.

In Secs. II through VII below the presently available data is surveyed, and in Secs. VIII through XVI theoretical ideas are discussed.

II. THERMAL STRUCTURE OBSERVATIONS

We discuss in this section only those features of the thermal structure of the Venus atmosphere that are particularly relevant to understanding the circulation. For additional information, the reader is referred to an accompanying
CIRCULATION OF THE VENUS ATMOSPHERE

chapter by Crisp and Titov, which presents a detailed summary of what is known concerning thermal structure and energy balance.

Pioneer Venus entry probes determined temperature profiles at four widely separated locations on the planet, with an uncertainty of \( \pm 1 \) K (Seiff et al. 1980). The Sounder Probe entered approximately on the equator near the morning terminator at local time 7:38 a.m. The North Probe entered at 3:35 a.m. at about 60 deg latitude. The Day and Night probes entered at about 30 deg south latitude, at local times of 6:46 a.m. and 0:07 a.m. Potential temperature profiles from all four probes are displayed in Fig. 1, calculated from the temperature, pressure tables given by Seiff et al. (1980). These profiles expand the temperature scale, display the stability, and permit comparison of temperatures. They show horizontal temperature contrasts less than 10 deg, and they show that the stratification is similar at all probe locations. There have also been a dozen Russian Venera probes into the Venus atmosphere, and all show temperature profiles very similar to those of Fig. 1 (Avduevskiy et al. 1983). Figure 1 does not display Venera data because the accuracy of these measurements (5 K; Avduevsky et al. 1983) is not sufficient to give a reliable stratification.

The Pioneer Venus Orbiter (PVO) carried an infrared radiometer with bands selected to permit retrieval of temperature profiles within the middle atmosphere (above the clouds). Figure 2 shows a latitude, height cross section and a longitude, height cross section in solar-fixed coordinates (Schofield and Taylor 1983). There is very little longitudinal temperature variation near the 1 bar pressure level but at higher altitudes a diurnal variation occurs, with an amplitude of about 5 K at the 10 mb level. The semi-diurnal component is the largest. These temperatures are evidence of solar-induced tides, and are extremely valuable for constraining the modeling of the tides on Venus (see Sec. XI below). Infrared spectra of Venus were obtained from the Venera 15 orbiter in 1983 (Schäfer et al. 1987), and thermal structure conclusions are consistent with those from the 1979 PVO.

The Russian/French Vega balloon experiment returned a large amount of data about temperatures within the cloud region from 50 to 55 km elevation, confirming that the stability is extremely small (Crisp et al. 1990). These results are described in Sec. VI below.

The thermal structure of the Venus atmosphere consists of a stable middle atmosphere, above the 0.2 bar, 60 km level, overlying a deep, less stable troposphere. At elevations above 50 km, where pressures are less than 1 bar, latitude-height temperature cross sections can be constructed from radio occultation data and thermal emission data, and the nature of the cyclostrophic balance can be examined (see Sec. III below). Longitudinal variation of temperature at these levels shows evidence of tides, but better vertical resolution would be useful. In the deep atmosphere there is an important gap in information within the bottom scale height because all the Pioneer Venus probes experienced electrical anomalies which affected the thermal sensors. Above 12 km the Pioneer probes determined the stability of the atmosphere. Figure 1
Figure 1. Temperatures from the Pioneer Venus probes, relative to an adiabat. This "potential temperature" would be uniform in a neutrally stable layer. Thermodynamic properties of the gases were taken from Hilsenrath et al. (1960).

shows the four temperature profiles relative to a reference adiabat. There are two layers with low stability, one in the cloud deck between about 0.5 and 1 bar, and one between the 8 and 30 bar level. Between 1 and 8 bars and in the upper atmosphere the thermal structure is stable. In Sec. VII we shall examine the correlation of the stability structure with the wind field at the Pioneer Venus probe entry sites.

III. RADIATION BALANCE OBSERVATIONS

The height profiles of upward and downward solar fluxes were measured by the Pioneer Venus Sounder Probe from the surface to about 64 km elevation (Tomasko et al. 1980; see also the chapter by Crisp and Titov). From this data one can calculate the net flux and the profile of energy deposition in the atmosphere. Approximately 17 W m\(^{-2}\) are absorbed at the surface. This is 11% of the net global average solar absorption \(\bar{F} = 157\) W m\(^{-2}\), and 2.6% of the global average insolation, 655 W m\(^{-2}\). More than half of the solar
energy is absorbed in the upper cloud layer between 55 and 70 km elevation. Of this, approximately half is absorbed by an unknown ultraviolet absorber in the cloud particles, and most of the rest is absorbed by $\text{H}_2\text{SO}_4$ aerosols (Crisp 1986).

The net global solar absorption is balanced by thermal emission to space. A radiative time constant based on the average thermal cooling rate and the entire mass of the atmosphere is $t_R = c_p T/\sigma F \sim 100$ yr, for $c_p = 800$ J kg$^{-1}$, a surface pressure of 90 bar, a mean temperature of 600 K, and an acceleration of gravity of 8.7 m s$^{-2}$. On the other hand, a radiative time constant based on the atmospheric mass within and above the cloud deck is only a few days (Crisp 1989), consistent with the observations indicating the existence of solar-induced tides.

In spite of uncertainties in the details, the fundamental radiative drive for large-scale motions in the deep atmosphere is known. Horizontal temperature gradients are small, and therefore the chief contribution to horizontal variations of net radiative heating is the variation of the insolation with distance away from the subsolar point, which has been established by the Pioneer probe measurements reported by Tomasko et al. (1985).
IV. CLOUDS

The Venus clouds are discussed in the chapters by Esposito et al. and by Crisp and Titov. Here we treat only topics related to dynamics.

Is the Venus cloud deck coupled actively to the atmospheric motion field? A major portion of the solar energy absorption is in the cloud, but if the cloud position is determined by microphysical and chemical processes that are independent of the atmospheric dynamics, then the answer could be negative. For the purposes of dynamical investigations, the cloud would then be a prescribed location of certain radiative forcings. The answer to the question is not yet fully understood, but several remarks can be made.

Nowhere in the Venus clouds did the Pioneer probes indicate a mass loading exceeding 100 mg m\(^{-3}\) (Knollenberg and Hunten 1980), corresponding to a mass mixing ratio of less than 10\(^{-3}\), comparable to a fairly thick cirrus cloud on Earth. The latent heat of evaporation is an order of magnitude larger than the heat content of the condensate, but because the mixing ratio is small, a phase change can produce a temperature perturbation of at most a degree or two. In contrast, the large-scale latitudinal contrasts near cloud level are on the order of 10 K (Fig. 2). Latent heating is therefore small in the Venus clouds, and coupling to motions by this mechanism is not likely to be important except possibly for effects on local convection.

Esposito et al. (1983) discuss the microphysics of the Venus clouds, and show that droplet formation times near the cloud tops are probably on the same order as fallout times (based on a scale height). Both are on the order of several weeks. Large-scale vertical velocities have been self consistently estimated by Newman and Leovy (1992) in a tidal computation that includes adjustment of the mean state, and the mean flow advection time (vertical displacement equal to a scale height) is also a few weeks. Therefore it appears that the large-scale motion field should cause a major perturbation to the cloud position, but probably would not be the controlling factor. Indeed there is a latitudinal gradient of cloud-top properties, including a “polar collar” of bright material at near 60° latitude and a polar region of high infrared emission at still higher latitudes (Taylor et al. 1980) which corresponds to a depression of the cloud-top elevation. Thus the large-scale motion field does appear to alter the cloud-top position but only as a perturbing effect.

Near the cloud base at about 50 km elevation the situation is more complicated. Thermal emission at near-infrared wavelengths escapes to space through gaps in the CO\(_2\) spectrum and permits determination of the total cloud optical depth. Groundbased observers of the night side of the planet have been able to map opacity variations (Crisp et al. 1991), and the Jupiter-bound Galileo spacecraft obtained infrared images with spatial resolution better than 100 km (Carlson et al. 1991). The opacity variations, which almost certainly arise in the lowest few km of the cloud, are approximately 25%. Radiative fluxes are altered by opacity variations, and there is the possibility of feedback between dynamics and the cloud structure. The pattern of cloud
opacity variations is streaky at high latitudes and more patchy at low latitudes, similar to the ultraviolet features. This region of the atmosphere may contain important dynamical activity and is not well understood.

V. FLOW OBSERVATIONS: LARGE SCALE-FEATURES

Here we discuss flows at cloud level and below. For discussion of middle atmosphere dynamics, see the chapter by Lellouch et al. Figure 3 displays profiles of the zonal and meridional wind with latitude deduced from tracking small scale features in the blue or ultraviolet seen in spacecraft images. Figure CDP3C6F1 displays an image of Venus that shows a typical pattern of features. They are probably formed just below the level where the ultraviolet optical depth is unity, which is about 40 mb pressure and 70 km elevation (Kawahata et al. 1980). The depth may vary by a few km with latitude and with time, but probably not by much more than this because the gradient of optical depth is large; the particle scale height is approximately the same as that of the gas, or about 4.8 km (Kawahata et al. 1980). The observed velocity profiles, which are each averages over several days of data, vary from epoch to epoch (Rozso 90 et al. 1990). These profiles are not necessarily representative of a zonal mean, because only the sunlit side of the planet is observed and there may be a solar-fixed tide with zonal or meridional amplitude on the order of 10 m s\(^{-1}\) (Del Genio and Rozso 90; Newman and Leovy 92).

The meridional velocities of Fig. 3 are of particular interest because they might give an indication of the strength of the Hadley circulation at cloud-top level. But unfortunately the uncertainty about the zonal mean is particularly important in this case because the tide and the mean may well be of the same order. The calculations of Newman and Leovy predict that the zonal mean is considerably smaller than the velocities shown in Fig. 3. If this is correct, the nightside velocities are small or even equatorward. On the other hand Schinder et al. (1990) have argued that the global spiral shape of the ultraviolet cloud features is consistent with poleward drift with zonal mean meridional velocities approximately equal to the observed dayside values. Further work with the cloud patterns may yield information on the important question of the relative strength of tidal and Hadley circulations. But there is no doubt that a tidal component exists in the cloud-tracking wind velocities. Figure 4 displays the latitude and longitude variation of the time-averaged zonal and meridional wind from Pioneer data (Del Genio and Rozso 90), and compares it with the predictions of Newman and Leovy. Other observation periods give similar results (Limaye 87; Del Genio and Rozso 90).

Another strong flow component at cloud top is the traveling four-day wave that produces the global pattern shaped like a “Y” or a “Ψ” rotated to the left to lie on its side. Early observers noted the motion of this feature (Boyer and Camichel 61; Boyer and Guérin 69). The structure of the albedo pattern became clear in Mariner 10 images (Belton et al. 76a) With Pioneer Venus and Galileo data it became possible to measure perturbations
associated with the feature (Del Genio and Rossow 1990; Rossow et al. 1990; Smith et al. 1993). The zonal wind oscillation at cloud top level is variable in magnitude but reaches 10 m s$^{-1}$. Meridional winds are smaller. This is consistent with early suggestions that the feature is a wave of Kelvin type
Figure 4. Time averaged velocity fields in a solar-fixed reference frame, at cloud top. Zonal wind is on the left and poleward wind is on the right. Panels (A) and (B) display calculated results from Newman and Leovy (1992), and panels (C) and (D) give the zonal and meridional components from measurements by Del Genio and Rossow (1990).

(see, e.g., Belton et al. 1976b). Theoretical calculation of wave structure was carried out by Covey and Schubert (1982), who showed that more than one free mode exists with the observed frequency. Smith et al. (1993), show that the Kelvin mode alone can produce the Y pattern and the flows associated with the waves as observed during the Galileo flyby. They also speculate that the wave is generated by a radiative-dynamical feedback at cloud base. It is important to learn the forcing and dissipation mechanisms for the four-day wave, because its momentum transfers depend on them and it appears to be a major dynamical phenomenon. Del Genio and Rossow and Rossow et al. also detected a disturbance with a five-day period, and identify it as a Rossby mode.

The solar tide, the four-day wave and the mean flow are the major components of the Venus large-scale flow near the cloud top. In fact, there is remarkably little else at this level. After removing the solar-fixed component and the four-day wave from the 1990 Galileo imaging data, Toigo et al. (1994) find an upper limit of about 4 m s$^{-1}$ on the residuals, which would include all synoptic and mesoscale time-dependent eddy and wave activity. Del Genio
and Rosswow (1990), analyzing PVO images, find transient waves at some epochs but also with amplitudes of only a few meters per second.

Flow at deeper levels has been obtained by tracking entry probes. Unfortunately the Pioneer and Venera probes, because of their aerodynamic properties, acquired the horizontal velocity of the atmosphere to within a few m s\(^{-1}\) only beneath the 60 km level, and therefore there is a gap of about a scale height between the top of the vertical profile from the probes and the ultraviolet feature tracking level. At altitudes less than 55 km the probes give the horizontal wind to within about 1 m s\(^{-1}\), with an altitude resolution of 1 km deep in the atmosphere and a few km near 55 km elevation (Counselman et al. 1980). Figure 5 displays the four Pioneer Venus entry probe profiles of zonal velocity, from Counselman et al., and also the locations of the entry points. These profiles show that the spin of the atmosphere is a global phenomenon, and that the spin begins to build up in the top half of the bottom scale height and continues to amplify right through the low and middle portions of the cloud deck.

Meridional velocities are also displayed in Fig. 5. There is no clear evidence for a Hadley circulation, but we shall see in Sec. VII below that the expected amplitude beneath the cloud is so small that none should be apparent in this data even if it exists as an important dynamical flow component.

Magalhães and Seiff (1995) have carefully looked for eddy structures in measurements of temperature from the atmospheric structure experiment on each of the Pioneer Venus entry probes using a spatial filtering technique. The filter separates structure with vertical scales of \(\leq 10\) km (eddy structures) from larger-scale structures (background structures). Between approximately 42 km to 60 km altitude, all the probes show eddy structure with amplitudes of about 1 K, well above the uncertainty level of the measurements, with a well-defined vertical wavelength of about 6 km. This structure is quite well correlated between the three small probes and some correlation with the fourth Large Probe is evident as well. Correlating the temperature fluctuations with velocity fluctuations is more difficult and has not yet been done, because the velocity fluctuations, of order 2 m s\(^{-1}\), are the same order as the velocity measurement errors, about 1 m s\(^{-1}\) (Counselman et al. 1980). The correlation in temperature between the probes, which were well separated in horizontal spatial position, suggests a dynamical phenomenon with a large horizontal scale, such as a planetary wave or atmospheric tide.

The two Vega balloons sampled the large-scale flow in 1985 at about 53 km elevation and at latitudes of about \(\pm 7\) deg (see description in Sec. VI). The east-west velocities were about 65 m s\(^{-1}\), confirming measurements from descent probes (Crisp et al. 1990). The north–south velocity showed a poleward drift by the southern probe with an average velocity of about 2.5 m s\(^{-1}\), and a much smaller northward velocity for the other. There is also a large scale acceleration pattern that may be due to a solar-fixed or topographically induced flow component.

It has recently been demonstrated that tracking of the near infrared hot
Also indicated are the descent locations of Russian Venus probes.

Locations of entry probes, "D," and "N" and "N" are destinations for two probes that entered the southern hemisphere.

Figure 5. (Left) Zonal and meridional velocity profiles from Pioneer Venus probe tracking (Conzemius et al., 1980). (Right)
spots, described in Sec. IV, gives velocities in the lower cloud region (Crisp et al. 1991; Carlson et al. 1991). In addition, Galileo imaging at 1 μm wavelength showed features that moved with about the same velocity (Belton et al. 1991). Thus it appears that near-infrared remote sensing can provide dynamical data from deep cloud layers, supplementing the traditional ultraviolet dynamical data from the cloud tops. At present the deeper data is sparse and not of high accuracy, but it offers promise of statistically useful results from a region of the Venus atmosphere that is not easily accessible by other means (see the chapter by Taylor et al.).

Newman et al. (1984) and Walterscheid et al. (1985) have used Pioneer Venus radio occultation data to estimate the latitude and height dependence of the zonal mean thermal structure for pressures between about 2 mb and 1.4 bar, and then integrated the cyclostrophic balance Eq. (4) to determine winds. They assumed a certain latitudinal profile of velocity at the 1.4 bar level. The dominant feature in their deduced zonal wind cross sections is a jet about one scale height above the ultraviolet cloud level with a maximum of about 120 m s⁻¹ at about 50 deg latitude. At higher levels the wind decreases. There is some evidence that the position and strength of the jet vary in time.

In the deep atmosphere, probe data suggests that the latitudinal velocity profile is not far from solid body rotation at each height (Schubert [1983] examines Pioneer data; Venera probes are consistent), although the sampling is sparse. Thus the evidence suggests that a transition takes place near 40 km elevation, above which the latitudinal profile of angular velocity develops a jet-like structure that is more pronounced with increasing height, until the core of the jet is reached just above the ultraviolet cloud-top level.

Another dynamical region of interest is at high latitudes, in either hemisphere. Pioneer Venus infrared measurements show peculiar double hot spots at approximately 80 deg northern latitude, located 180 deg apart in longitude (Taylor et al. 1980). These authors attribute the structure to depressions in the cloud top elevation. Ultraviolet images of high latitudes appear symmetric in longitude and do not show the double structure. It is possible that there is a disturbance in the middle cloud region, too deep to be detected in the ultraviolet images. Stability studies of the circumpolar jet show that it may be unstable to wavenumber 2 (Elson 1982; Young et al. 1984; Michelangeli et al. 1987).

VI. FLOW OBSERVATIONS: SMALL SCALES

Ultraviolet images of the Venus cloud tops show blotches that are roughly circular, streaks that spiral toward the poles, and occasionally long linear features (Belton et al. 1976a; Rosson et al. 1980). Small features are typically between 200 and 1000 km in size. Interpretation is hampered because the ultraviolet absorber is unknown, and the process that causes darkening or lightening is not known. The albedo modification probably involves vertical displacement or vertical velocity. Del Genio and Rossow (1990) argue from
the physics of Kelvin waves that positive vertical velocity corresponds to the
dark region of the large scale “Y,” but the exact phasing is uncertain, and small-
scale features may behave differently from large ones. Schinder et al. (1990)
showed that the pattern of streaks spiraling toward the pole can be explained
by poleward advection and shearing of cloud patches formed at low latitudes,
using the observed large-scale pattern of zonal and meridional velocity. Toigo
et al. (1994), using high-resolution Galileo feature tracking data, show that
small features have a lifetime of about two days. The Galileo data sequence
was designed to search for small-scale high-frequency wave activity (Belton et
al. 1991) and none was detected. It has been speculated (Belton et al. 1976b;
Baker and Schubert 1992) that the isotropic blotches in the cloud top are
manifestations of convection, but this is uncertain. Toigo et al. 1994 argue that
the features are mesoscale dynamical cells driven by local radiative heating
inhomogeneities caused by variable amounts of the mysterious ultraviolet
absorber, with dynamical feedback reinforcing the absorber inhomogeneities.
Contrasts in the ultraviolet are on the order of a few percent, enough to produce
dynamically important heating.

Two balloons with tethered instrument packages were placed in the Venus
atmosphere in 1985 by the Russian-French Vega mission. Meteorological
measurements are reported by Crisp et al. (1990). The balloons were inserted
four days apart and each drifted for approximately two days, covering more
than 100 deg of longitude, at an elevation between 50 and 55 km. One was
about 7 deg north of the equator and the other the same distance south. During
their vertical excursions of 2 or 3 km they measured temperature gradients
very close to adiabatic, consistent with the low stability zone apparent in Fig. 1
near the 600 mb level. Remarkably, the absolute temperatures measured by
the two differed by 6.5 K, in spite of their small difference in latitude. If
correct, this measurement indicates surprisingly large temperature variations.
Vertical motions inferred from pressure variations are a few meters per second,
and correlations of velocity with temperature imply a vertical heat flux on the
order of 30 W m⁻² with one burst an order of magnitude larger. These results
are consistent with thermally driven convection due to the high opacity of the
cloud to infrared radiation. There is a suggestion that activity is correlated
with the position of topographic relief on the planet’s surface. Young et al.
(1987, 1994) show that topographically excited stationary waves can penetrate
to this height from the surface.

Schinder et al. (1990) show that the regions just above and just below the
convecting layer within the cloud can act as ducts for horizontally propagating
gravity waves. The strong wind shear above the low stability zone and the
high static stability below it produce the ducts. Leroy and Ingersoll (1995a)
show that waves do not propagate a large distance before being absorbed,
however. Radio occultations show scintillations both above and below the
convective layer that may be due to turbulence or trapped gravity waves (Woo
et al. 1980; Leroy and Ingersoll 1995b).

The best observational evidence for the existence of small scale inter-
nal waves comes from two recent studies, one involving radio occultations obtained by the Magellan spacecraft (Hinson and Jenkins 1995), and the other an analysis of Pioneer Venus entry probe data (Magalhães and Seiff 1995). Hinson and Jenkins show that near and above the middle clouds, small-scale oscillations in retrieved temperature profiles and scintillations in received signal intensity are consistent with a spectrum of vertically propagating radiatively damped gravity waves. There appears to be one wave that predominates, having a vertical wavelength between 2 to 3 km and wave amplitude of about 4 K at 65 km altitude. The wave is nearly stationary with respect to the surface or the Sun. Computed wave attenuation due to radiative damping suggests that the wave contributes to wave drag on the mean zonal winds with a magnitude of about 0.4 m s$^{-1}$ per day. Convective activity in the neutrally stable layer of the middle cloud deck would be unlikely to produce a wave stationary with respect to either the surface or Sun. Possible wave sources might be a stationary patch of turbulence in the middle cloud deck due to flow over topography (Gierasch 1987), bow waves set up by a subsolar disturbance (Belton et al. 1976), or topographically generated waves (Young et al. 1987, 1994).

Below 40 to 42 km altitude, the Large, North, and Day probes show small-scale structures with amplitudes of about 0.2 to 0.5 K and a well-defined vertical wavelength of 4 to 8 km (Magalhães and Seiff 1995). No correlation between the probes is evident for the structures at these altitudes, implying that the fluctuations are probably due to small-scale or mesoscale activity. Magalhães and Seiff (1995) find that the phase and amplitude variation of the temperature fluctuations observed below 42 km altitude are well accounted for by the WKB solution for a single linear internal gravity wave mode propagating through the observed background structure. The large variation in static stability and zonal wind with altitude over the range of the observations leads to a strong sensitivity of the phase variation of the internal gravity wave to the assumed zonal phase speed and horizontal aspect ratio of the wave (defined as the ratio of the meridional wavenumber to zonal wavenumber), thus providing a constraint on these properties. A three-dimensional internal gravity wave with a zonal phase speed of approximately 30 m s$^{-1}$ and horizontal aspect ratio of about 2 accounts best for the Large Probe observations. The North Probe observations are well reproduced by a wave with zonal phase speed 0 to 15 m s$^{-1}$ and aspect ratio near 5. Waves generated near the surface either by convection or topography would be expected to have phase speeds near zero. However, from geometry, large zonal phase speeds can be associated with waves propagating mostly in the meridional direction even though the phase speed in the direction of propagation along the total wave vector is small. As will be discussed later in connection with convectively driven waves, such waves would not be expected to have significant zonal momentum transport. A wave generated higher in the atmosphere would likely propagate slowly with respect to the mean wind at the altitude of wave generation, and could thereby have a relatively large phase speed with respect to the surface.
Another indication of small or mesoscale activity on Venus is the diverse
directionality of wind streaks observed by the Magellan radar. These are
features typically a few km in size that are probably formed in the planetary
boundary layer (Greeley et al. 1992, 1995). The north–south component of
the direction is somewhat more often equatorward than poleward (Greeley
et al. 1992, 1994), suggesting that a Hadley circulation exists, but a large-
scale pattern is weak, if it exists at all. This is to be contrasted with the
large (500 km) parabolic dark markings that occur (Campbell et al. 1992)
and invariably point toward the east. These features are probably due to
large impacts, with the open end of the parabola representing the downwind
direction in the upper atmosphere. They suggest that the atmospheric rotation
existed in the past in the same direction as at present.

VII. FLOW OBSERVATIONS: DISCUSSION

Figure 7 displays profiles of the squared Brunt frequency \( N^2 = g/T[(dT/dz)\
-(dT/dz_{ad})] \) and the squared shear, \( S^2 = (\partial u/\partial z)^2 + (\partial v/\partial z)^2 \), calculated
from the Pioneer Venus probe profiles. Here \( g \) is the acceleration of gravity,
\( T \) is temperature, \( z \) is height, and \( u \) and \( v \) are the eastward and northward
velocity components. There appears to be a positive correlation between the
thermal stratification and the shear from 20 km to 60 km elevation. There are
two layers, one from about 50 to 55 km and one from 20 to about 35 km, where
both the stability and the shear are relatively small. Between 10 and 20 km
elevation there is a maximum in the shear, and there is some indication that it
is reflected in the stratification, but unfortunately the absence of Pioneer probe
temperature data below 12 km makes it impossible to evaluate the stability
(which requires a gradient) below about 15 km.

The Richardson number, \( Ri = N^2/S^2 \), can be evaluated from the ratio
of the stability and the shear and is displayed in Fig. 7. It is the ratio of two
noisy quantities and is displayed as a scatter plot to help the reader assess
its significance. The stability and shear profiles were smoothed again before
forming the ratio. Figure 7 shows that there are two layers where \( Ri \) is of
order unity, each about a scale height thick, one centered near 24 km and
one near 54 km. \( Ri \) is on the order of 10 in a deep layer between 30 and
50 km. There is some indication that \( Ri \) increases below 20 km, but the lack
of thermal data makes this very uncertain. There is also the possibility that
near the base of the atmosphere there is another layer with small stability.
Radiative heating usually destabilizes the boundary layer of an atmosphere
when any solar heating reaches the surface (Goody and Yung 1989). Thus
there are at least two, and possibly three, layers in the Venus atmosphere
with small \( Ri \) (of order unity or less) sandwiched between more stable layers.
The Richardson number is an important stability parameter, and in addition,
Allison et al. (1994) have shown that it can be used to place constraints on the
latitudinal shear at low latitudes. More extensive and accurate determinations
would be valuable.
Figure 6. Shear and static stability profiles from the Pioneer Venus probes. The data was smoothed with a 5.5 km sliding box average before taking derivatives.

In Fig. 8 we have plotted the Pioneer Venus meridional velocities compared with two amplitude envelopes. The kinetic energy density is \( \rho V^2/2 \sim \text{const.} \), where \( V \) is the magnitude of the full three-dimensional velocity and \( \rho \) is the gas density. Vertically propagating waves tend to conserve their kinetic energy density and therefore display velocity components that are approximately proportional to \( \rho^{-1/2} \). Another characteristic behavior would arise if turbulence existed at all levels and dissipated energy per unit volume at approximately equal rates at all heights. In isotropic turbulence the kinetic energy dissipation rate is proportional to \( \rho V^3 \), leading to a velocity perturbation that would vary as \( \rho^{-1/3} \). Figure 8 shows no preference for one of these particular behaviors over the other, but it convincingly shows a height dependence of the velocity amplitude that is inversely related to the density. The velocity profiles do not show any obvious correlations between different probes, nor do they show vertical coherence that would indicate long trains of vertically propagating waves. The profiles, especially beneath 30 km elevation, must be viewed with caution because of \( \pm 1 \text{ m s}^{-1} \) uncertainties in Pioneer Venus velocities. Improved data of this kind is needed to answer questions about waves and turbulence, and samplings at more closely spaced horizontal stations are needed to establish the horizontal coherence scale of the motions.

It was first pointed out by Leovy (1973) that the large-scale latitudinal
force balance in the Venus atmosphere is cyclostrophic. This balance is the analog of the geostrophic balance that obtains on rapidly rotating planets, with the centrifugal acceleration replacing the Coriolis acceleration (Holton 1992). The vertical force balance on large scales is hydrostatic. Let \( u_a \) be the absolute zonal flow velocity relative to a nonrotating frame, \( a \) be the planetary radius, \( \phi \) be latitude and \( z \) be height. The cyclostrophic analog of the thermal wind equation of meteorology (Holton 1992) is

\[
H \frac{\partial}{\partial z} u_a^2 = - \cot \phi \left[ \frac{\partial}{\partial \phi} (RT) \right]_{\rho \text{constant}}
\]

where the derivative with respect to latitude is at constant pressure, and \( H \) is the scale height, \( H = RT/g \), with \( R \) the gas constant. In the deep atmosphere Eq. (1) can be used to estimate horizontal temperature gradients from observed wind shears. Near and above cloud top, it has been used to infer wind shear from measured temperatures, as discussed in Sec. V under the topic of radio occultation data.
Figure 8. Meridional velocity profiles for the Pioneer Venus probes compared against two envelopes.

It is interesting to compare the Pioneer Venus probe measurements with simple dynamical scaling estimates based on various extreme assumptions. One case is the meridional circulation that just balances the latitudinal gradient of solar heating. The vertical velocity is obtained by assuming that adiabatic cooling balances the solar heating at low latitudes (Holton 1992):

$$\frac{N^2}{g} w = \frac{Q_R}{\rho c_p}.$$  (2)

The radiative heating $Q_R$ can be evaluated by using the solar flux, because it is the latitudinal gradient of the solar flux that gives the equator to pole radiative heating differential, to a first approximation. Solar fluxes from Tomasko et al. (1980) were used. Figure 9 displays a smoothed profile of the Brunt frequency that was adopted. The horizontal velocity is estimated from the vertical velocity by multiplying by the ratio of the planet's radius to the local scale height.

An exception occurs where the Brunt frequency is very small because adiabatic cooling cannot balance the solar heating. In that case, the estimate
is replaced by a nonlinear one based on the assumptions (Gierasch et al. 1970)

\[
\frac{v \Delta T}{a} = \frac{Q_R}{\rho c_p}
\]

\[
v^2 = H g \frac{\Delta T}{T}.
\]  

(3)

The horizontal temperature contrast \(\Delta T\) can be estimated from the cyclostrophic balance result Eq. (1). Equations (1), (2) and (3) give

\[
v = \frac{agQ_R}{HN^2 \rho c_p T} \quad \text{for} \quad N^2 > \frac{(agQ_R)^{2/3}}{H^2 \rho c_p T}
\]

\[
v^3 = aH g \frac{Q_R}{\rho c_p T} \quad N^2 < \frac{(agQ_R)^{2/3}}{H^2 \rho c_p T}
\]  

(4)

The meridional velocity is displayed in Fig. 10. Near 15 and 40 km, where the assumed stratification is large, the velocity is on the order of a few cm s\(^{-1}\). At sub-cloud levels it is only 2 or 3 m s\(^{-1}\) even at those heights where the assumed stratification is very small. Furthermore, these estimates assume that the entire equator to pole solar heating imbalance is carried by a Hadley cell, and therefore they are upper limits. At cloud top levels the estimates give about 10 m s\(^{-1}\), which is comparable to calculated (and observed) tidal amplitudes.

Another interesting index for comparison is the free convection velocity of mixing length theory, for a mixing length equal to a scale height. This velocity is based on the assumptions (Priestley 1959)

\[
WT' = \frac{F}{\rho c_p} \frac{W^2}{H} = g \frac{T'}{T}
\]  

(5)
Figure 10. Scaling estimates of the amplitudes of velocity and temperature contrasts under different simple assumptions. Convective velocities are from mixing length theory, based on a mixing length of a scale height and the energy flux profile from Fig. 4. Hadley velocities are based on the same flux profile and the Brunt frequency profile from Fig. 9. The Hadley curve is in segments because two different kinds of estimates are made, depending on the stratification at different heights, as discussed in the text. Wave amplitudes are scaled inversely by the square root of the density, and arbitrarily set to 10 m s$^{-1}$ at 60 km elevation. The cyclostrophic temperature contrast is of global scale, and is based on Eq. (1) and the zonal wind profile from Fig. 9. The other temperature amplitude estimates are discussed in the text. The dashed curves show theoretical boundary layer slope wind estimates (Sec. XIV).

where $F$ is the forcing heat flux, in this case due to solar radiation, $W$ the velocity, $H$ is the mixing length, taken as the local scale height, and $T'$ the temperature fluctuation amplitude. Solving for $W$ and $T'$ gives

$$W^3 = \frac{R H g F}{c_p \rho} \frac{T'}{T} = \frac{W^2}{H g}$$  \hspace{1cm} (6)

after using $H = RT/g$. All three velocity components are assumed to be the same magnitude in this turbulent convection scaling estimate. The profile of $W$ with height is displayed in Fig. 10 as the “convective” curve, with $F$
evaluated from the estimated profile of global mean solar flux (Tomasko et al. 1980). This velocity scale might be relevant near 55 km, near 25 km, and possibly near the surface, where the stratification is weak or negative.

Figure 10 also displays a velocity envelope proportional to $\rho^{-1/2}$ like that indicated in Fig. 8, which might be a result of vertical wave propagation. The surface amplitude was set at 0.5 m s\(^{-1}\), which gives fairly good agreement with meridional velocity fluctuations of about 10 m s\(^{-1}\) at 60 km elevation in the data (Fig. 8). Figure 10 also indicates wave amplitude predictions (dashed line) for slope-induced waves or diurnal slope winds, which would exist only over elevated or sloping terrain (Sec. XV).

Another kind of motion that might well exist in the stratified regions of the Venus lower atmosphere is two dimensional eddy activity. The source could be instability of the horizontal shear. These motions would not propagate vertically. Their amplitude would be given by the strength of the horizontal shear in the zonal mean wind, which is unknown because the sampling density in the lower atmosphere is so small. Layers of quasi-two-dimensional turbulence can exist if the Richardson number is greater than unity (Herring and Météis 1989).

Figure 10 also presents temperature amplitude estimates for various flow components. The latitudinal temperature contrast associated with cyclostrophic balance is calculated from Eq. (4), using the smoothed zonal velocity profile shown in Fig. 10 and assuming a solid body rotational profile at each elevation (constant pressure surface). The mean temperature profile from the Pioneer Venus Sounder probe (Seiff et al. 1980) is used to evaluate the scale height and the temperature where they occur as coefficients. The heat equation, continuity equation and dispersion relation for gravity waves (Holton 1992) give an estimate of the temperature oscillation associated with internal waves

$$\frac{T'}{T} = V \frac{N}{g}$$

(7)

where $V$ is the horizontal velocity perturbation associated with the wave. Finally, the convective temperature fluctuation, from Eq. (6), is also displayed. The latter is too small to be responsible for the measured probe temperature fluctuations. The wave estimate, although a bit small, is consistent with the observations. Also consistent with the observed fluctuations would be two-dimensional eddy activity with horizontal advection of the latitudinal temperature contrasts associated with cyclostrophic balance.

**VIII. MAINTENANCE OF THE CIRCULATION: GENERAL DISCUSSION**

The angular momentum per unit mass is

$$M = \Omega r^2 \cos^2 \phi + ur \cos \phi$$

(8)
where $r$ is the spherical radius, $\phi$ is latitude, and $u$ is the zonal (tangential) velocity. The equations of motion give

$$\frac{\partial}{\partial t} (\rho M) + \nabla \cdot (\rho v M) + \frac{\partial p}{\partial \lambda} = \nabla \cdot (\tau \times r)$$  \hspace{1cm} (9)

where $\lambda$ is longitude, $\hat{z}$ is the unit vector in the pole direction, and $\tau$ is the viscous stress tensor. If Eq. (9) is averaged in longitude the pressure gradient term drops out. The remaining terms express the effects of advection and of friction on the longitudinally averaged angular momentum distribution. The continuity equation is

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho u) = 0.$$  \hspace{1cm} (10)

For axisymmetric flow with negligible friction Eqs. (9) and (10) give

$$\frac{\partial M}{\partial t} + u \cdot \nabla M = 0.$$  \hspace{1cm} (11)

Thus under the influence of axisymmetric advection, fluid rings conserve angular momentum as they move in height and latitude, and there is never an extremum of angular momentum along a particle trajectory. The highest angular momentum that a fluid ring can acquire from the planet’s surface is that at the equator, and this is the maximum that axisymmetric advection of fluid rings can produce in the atmosphere (Hide 1969). Mid-latitude jets can be formed by poleward drift of material, because the angular velocity and the linear velocity will increase as the fluid moves closer to the rotation axis.

On Venus the angular momentum per unit mass of the atmosphere has an internal maximum at low latitudes and at an elevation of approximately 70 km. The angular velocity at this location is approximately 50 times that of the solid planet. Hide’s theorem shows that nonaxisymmetric motions must be at play, and must pump momentum toward this location. Specifically, Eq. (9) shows that there must be convergence of the angular momentum flux $F_M = \rho v M$ toward the location of the maximum of the angular momentum per unit mass. The key terms in producing these are the velocity correlation, or Reynolds’ stress terms

$$\overline{F_M} = \rho r \cos \phi \left( \overline{u'v'}, \overline{u'u'} \right)$$  \hspace{1cm} (12)

where $u'$ and $w'$ are the meridional and vertical components of the eddy (nonaxisymmetric) portion of the velocity field.

For conceptual simplicity, the hypotheses about the maintenance of the rotation of the Venus atmosphere can be divided into two classes, depending on whether the vertical or meridional Reynolds’ stress is most important. Under the first category a vertical flux of angular momentum by eddies at low latitudes drives the circulation. The key balance would then be vertical pumping of angular momentum by eddies, balanced by dissipative processes
that might involve other eddies, turbulence or small-scale friction. The eddies that produce the vertical pumping might be caused by solar heating (tides), by convectively driven gravity waves, or other processes. Under the second class of hypotheses the meridional flux of eddy angular momentum is most important. In this case angular momentum is pumped horizontally into low latitudes from high latitudes. At high latitudes angular velocity is generated by axisymmetric poleward drift in a Hadley circulation. The vertical balance of angular momentum does not involve eddies under this hypothesis. A Hadley circulation generally pumps angular momentum upward because the angular momentum per unit mass is larger at low latitudes (where the Hadley circulation is rising) than at high latitudes, and the vertical balance can be maintained by an upward angular momentum transport by the Hadley cell, offsetting downward momentum flux by dissipative loss, friction or turbulence.

These angular momentum flux requirements can be stated quantitatively by integrating Eq. (9) over different portions of the atmosphere. The rate of change of the net angular momentum of a layer of atmosphere from radius \( r_1 \) upward to infinity is given by

\[
\frac{\partial}{\partial t} \int_{r_1}^{\infty} \int_{-\pi/2}^{\pi/2} \int_{0}^{2\pi} \rho M r^2 \cos \phi \, d\lambda \, d\phi \, dr \\
= \left[ \int_{-\pi/2}^{\pi/2} \int_{0}^{2\pi} \rho \omega M r^2 \cos \phi \, d\lambda \, d\phi - \int_{-\pi/2}^{\pi/2} \int_{0}^{2\pi} \tau_{r\lambda} r^3 \cos^2 \phi \, d\lambda \, d\phi \right]_{r=r_1}
\]

(13)

where \( \tau_{r\lambda} \) is the \( r, \lambda \) component of the viscous stress tensor. For a steady state the radial flux of angular momentum by the flow must offset viscous drag. The radial flux of angular momentum can arise either by a large-scale axisymmetric Hadley circulation, because \( M \) is largest at low latitudes where \( \omega \) is positive, or it can arise because of a correlation of \( M \) and \( \omega \) associated with eddies. One or the other (or both) must exist.

In reality it is probably not viscous dissipation that must be overcome by organized angular momentum fluxes, but some combination of small-scale turbulence and wave drag. We shall elaborate on this point in Sec. IX. For simplicity, all “dissipative” mechanisms can be incorporated into \( \tau \) for our purposes in this section.

Integrating over the entire depth of the atmosphere and over a latitude range from \(-\phi_1\) to \(\phi_1\) gives another important statement,

\[
\frac{\partial}{\partial t} \int_{\phi_1}^{\phi_1} \int_{\phi_1}^{\phi_1} \int_{0}^{2\pi} \rho M r^2 \cos \phi \, d\lambda \, d\phi \, dr \\
= \left[ \int_{\phi_1}^{\phi_1} \int_{0}^{2\pi} (\rho \omega M - \tau_{r\lambda} r \cos \phi) r^2 \cos \phi \, d\lambda \, d\phi \right]_{r=r_1} \\
+ \left[ \int_{\phi_1}^{\phi_1} \int_{0}^{2\pi} (\rho \omega M - \tau_{\phi\lambda} r \cos \phi) r \cos \phi \, d\lambda \, dr \right]_{\phi=\phi_1}^{\phi=-\phi_1}
\]

(14)
There can be drag at the irregular bottom surface. In writing Eq. (14) we account for this by integrating only from \( r_{0+} \) to \( \infty \), where \( r_{0+} \) is high enough above the mean surface position \( r_0 \) not to intercept the topography. The first term on the right-hand side of Eq. (14) would normally be expected to act as a drag (although it has been suggested that wave propagation from near-surface layers might carry net momentum upward; see Sec. XII below). In the second term on the right latitudinal fluxes into the volume are evaluated. The viscous stress term here is expected to be small, leaving the transport of momentum by latitudinal motions as the dominant term. This can be decomposed into the transport by the mean Hadley circulation and the transport by eddies. One expects the Hadley circulation to transport angular momentum poleward, because the flow is poleward at high levels where the spin rate is greatest. Therefore the Hadley circulation will act as a net drag on low latitude flow, and a balance will require that eddies transport angular momentum equatorward. The only alternative is that the near-surface term at \( r_{0+} \) provides sufficient acceleration to offset the Hadley circulation.

Note that the vertical flux of angular momentum can be either by eddies or by the mean Hadley circulation, but that in the latitudinal direction the Hadley circulation is always expected to move angular momentum out of low latitude regions. Thus the Hadley circulation can act to assist the equatorial superrotation by producing an upward momentum flux, but is always expected to hinder it horizontally by moving angular momentum out of low latitudes.

The required magnitude of the acceleration by eddy momentum flux convergence can be estimated by evaluating the Hadley circulation accelerations, using Eq. (4) to estimate the Hadley circulation strength. In fact Eq. (4) gives an upper limit, because eddies can transport part of the equator–pole heat flux. For example, Newman and Leovy (1992) find that the Hadley circulation is very much suppressed at cloud-top level at low latitudes by the effects of tidal waves. Nevertheless, estimates based on Eq. (4) are probably correct in order of magnitude, and useful if viewed with caution. The acceleration produced by poleward flow is

\[
\frac{\partial u}{\partial t} = -\frac{v}{r \cos \phi} \frac{\partial}{\partial \phi} (u \cos \phi) = \frac{vu}{r}
\]

where it is assumed that the zonal flow, on a global mean, is not far from uniform rotation at each height. Values of the meridional velocity can be read from Fig. 10, and the zonal wind from Fig. 5. For example, at heights of 40 km and 70 km, Eq. (15) gives estimates of 1 cm s\(^{-1}\) day\(^{-1}\) and 10 m s\(^{-1}\) day\(^{-1}\), respectively. Near the equator it would be more appropriate to use \( w \, du/dz \) to make this estimate, and the result would be approximately the same because \( w \) and \( v \) in the Hadley circulation are related by continuity. Eddy strengths can then be estimated. For example, if horizontal momentum transports are to be important

\[
\frac{v' u'}{r} \sim \frac{vu}{r}
\]

(16)
Assuming that the two eddy components are the same magnitude gives \( v' \) equal to the geometric mean of the Hadley meridional velocity and the zonal flow, or about 1 m s\(^{-1}\) at 40 km and 20 m s\(^{-1}\) at 70 km. Similar estimates can be made for vertical eddy requirements.

IX. FRICTION AND WAVE DRAG

The spin of the Venus atmosphere is remarkable and puzzling because intuition tells us that in the absence of forcing, friction will bring the atmosphere into corotation with the solid planet. But the nature of "friction" is also unclear, as is its magnitude. It may be that the Venus atmospheric rotation is controlled by balances between different large-scale flow components, and that friction, in the sense of molecular viscosity or small-scale turbulent dissipation, is unimportant. This is one of the key questions to be resolved by future observations. In this section we shall make a few remarks about candidates for frictional mechanisms.

The kinematic molecular viscosity of CO\(_2\) is 0.07 cm\(^{-2}\) s\(^{-1}\) at STP. It is approximately proportional to \( T^{1/2} \rho^{-1} \), but even at high altitudes above the Venus clouds where the density is reduced by a factor of 100 from STP the time scale \( H^2/\nu \) is about 1000 yr. Molecular viscosity is therefore too small to have a direct influence on motions whose vertical scale of variation is a scale height.

Turbulence can act to produce an effective viscosity. At scales small enough so that anisotropies caused by stratification or mean shear become unimportant, turbulence is isotropic and kinetic energy migrates to small scales by a nonlinear cascade. This is the inertial, or Kolmogorov, regime. In this regime kinetic energy is removed by eventual viscous dissipation at very small scales. The fluid is brought toward its lowest energy state, which is rigid body rotation. Thus small-scale turbulence acts analogously to molecular viscosity and tends to erode angular velocity gradients. Prandtl's mixing length concepts can be used to estimate the effective viscosity if the eddy scales and velocities are known at the outer scale of the isotropic regime. The difficulty is that on Venus (and, incidentally, on Earth) the details of the velocity field and the stratification are not known well enough to describe the distribution of turbulent dissipation. Turbulence at larger scales can be anisotropic and does not necessarily act analogously to molecular viscosity. In fact, two dimensional turbulence is a candidate mechanism for producing equatorward momentum transports that help sustain the Venus atmospheric rotation (see Sec. XIII).

Wave drag is important in the Earth's atmosphere and is very likely the major deceleration mechanism that needs to be overcome on Venus. Waves of all scales are ubiquitous in atmospheres and are generated by a variety of sources, including convection, internal turbulent patches, flow over topography, and instabilities. The behavior of the Earth's atmosphere shows that the net effect of wave absorption is usually to brake the atmosphere (see, e.g.,
Holton 1982 or Andrews et al. 1987). Although many different sources exist with varied phase speeds, the presence of stationary irregularities at the bottom surface of an atmosphere produces a peak in the wave spectrum at zero horizontal velocity. Because absorption of waves accelerates the atmosphere toward the wave velocity, the net effect is to drag the atmosphere toward zero velocity.

There are notable exceptions when waves of one particular direction are favored and absorption tends to produce a net acceleration. The quasi-biennial oscillation in the Earth’s equatorial stratosphere is an example (Holton and Lindzen 1972), and the solar tides on Venus are another (see Sec. XI). But these are exceptions, and drag is the general rule. To evaluate the magnitude of wave drag on Venus we must determine the spectrum of waves, particularly in the lower atmosphere, and we must determine the absorption mechanisms. In the Earth’s mesosphere the dominant absorption mechanism is breaking of waves (Lindzen 1981; Holton 1982). At lower elevations radiative damping may play an important role in wave absorption (see, e.g., Holton and Lindzen 1972).

X. DIAGNOSTIC STUDIES

Hou (1984) and Hou and Goody (1985,1989) have deduced the momentum sources needed to maintain the Venus circulation for an assumed latitude-height distribution of zonal winds, temperatures and radiative heating. They assume that the vertical divergence of the eddy heat flux is smaller than the effect of vertical heat advection. Under these conditions the equations of motion can be inverted to give the momentum fluxes. These include the effects of both eddies and friction, but through most of the atmosphere friction is probably negligible relative to the eddies.

Two important conclusions about eddy Reynolds' stresses are probably valid in spite of any uncertainties about the mean flow and the neglect of eddy heat fluxes. One point is that over the height range between about 15 and 50 km, positive eddy accelerations (in the direction of the atmospheric spin) are necessary at low latitudes. This is the fundamental requirement for maintenance of the spin of the lower atmosphere. It is needed to offset the deceleration due to low latitude rising motion in a positive vertical shear. The other point is that from 50 to 80 km, also at low latitudes, an approximately offsetting pair of requirements exists, with deceleration at the highest levels, peaking between 70 and 80 km, and acceleration peaking near 65 km. This offsetting pair of requirements can be met by a downward flux of eddy momentum that is confined to the region near and just above the cloud tops. It is needed to balance the accelerations due to the Hadley circulation penetrating through the mean zonal flow vertical shear, that is positive just below the zonal flow maximum (at about 70 km) and negative above this.

This result suggests that two different eddy momentum transfer processes exist, and perhaps are accomplished by two different kinds of eddies. The distribution of zonal flow near and above the cloud tops requires a downward
momentum flux, which is opposite to the global requirement to maintain the atmospheric spin. Several scale heights deeper, the maintenance of the basic spin requires convergence of a momentum flux that is either upward from the surface or equatorward from high latitudes. It now seems likely that the solar tides accomplish the cloud-top requirements, as we shall discuss in the next section. The deep requirements remain unexplained.

XI. THE TIDES

Recent work has demonstrated convincingly that the solar tides produce a downward angular momentum flux just above the Venus clouds, which decelerates the stratosphere and accelerates the upper part of the cloud layer. These tides are global-scale motions forced by solar heating. They are internal gravity waves modified by rotational effects. The internal gravity wave vertical wavenumber is approximately \( k_V = (\omega / N) k_H \), where \( \omega \) is the frequency in the frame moving with mean flow and \( k_H \) is the (imposed) horizontal wavenumber. It turns out that for the semi-diurnal tidal component, \( k_V \) matches well the vertical half-width of the solar heating distribution (Crisp 1986), and as a result there is effective driving. Leovy (1987) argues that this is not a coincidence, as we discuss below.

Soon after the rotation of the Venus atmosphere was discovered, Schubert and Whitehead (1969) showed that a moving heat source can produce a mean flow in a laboratory annulus, and suggested that the flow on Venus is produced by the motion of the subsolar point. Theoretical work showed that phase shifts across the laboratory fluid produce Reynolds' stresses that drive the mean flow in the laboratory situation (see, e.g., Malkus 1970). Fels and Lindzen (1974) pointed out that thermally excited gravity waves also produce Reynolds' stresses. Wave phase shifts arise because of propagation, while the laboratory annulus develops phase lags because of viscous and conductive effects. Waves carry momentum away from the excitation region, causing a reaction on the excitation region that accelerates it in the opposite direction to that of the transmitted wave. Figure 11 illustrates the concept. In the case of Venus the atmospheric rotation carries the fluid past the subsolar point and a wave is established that is stationary with respect to the Sun, and moves "upstream" in the fluid, against the direction of the mean flow. The wave is excited by the absorption of sunlight primarily between 60 and 70 km elevation. Vertically, it propagates upward and downward from the principal excitation region. The upward component is absorbed by radiative damping a few scale heights higher, and produces a deceleration, which is balanced by acceleration of the cloud tops. The downward component is weaker and is not as important.

Pechman and Ingersoll (1984) calculated the structure of both the diurnal (zonal wavenumber 1) and semidiurnal tidal waves (wavenumber 2) and showed that the semidiurnal tidal component agreed well with Pioneer Venus observations of the stratospheric temperature (see Fig. 2). A series of increas-
Figure 11. Schematic of tidal waves and accelerations. Solar motion is eastward but slow and most of the velocity of the heat source relative to the fluid is due to the westward motion of the fluid. Momentum is deposited where waves are absorbed, and a reaction force is produced where waves are generated. In practice, the details of dissipation, heating and wave propagation characteristics are all important in determining the three dimensional distribution of accelerations. On Venus, the downward momentum flux by tidal waves is smaller than the upward flux.

Innovatively detailed theoretical treatments of the tides followed (Fels 1986; Leovy and Baker 1987; Hou et al. 1990; Newman and Leovy 1992). Newman and Leovy allow the mean flow to evolve, including the Hadley circulation, under the influence of both the diurnal and semidiurnal tidal components. At the base of the computational model, near 40 km elevation, they impose a mean flow as observed. They find excellent agreement with the latitude and height dependence of the mean zonal wind above the cloud tops that was deduced from radio occultations and with the latitude and longitude dependence of cloud tracked winds (Fig. 4). Leovy (1987) points out that the tidal forcing will tend to establish a value of the mean flow near \( u = N h \), where \( N \) is the Brunt frequency and \( h \) is the thickness of the heated forcing layer (approximately two scale heights), because this mean flow speed produces a good match of the tidal vertical wavelength with the scale of the forcing. The vertical wavelength of a low frequency internal gravity wave is smaller than the horizontal scale by the factor \( \omega/N \), where \( \omega \) is the Doppler shifted frequency in the frame moving with the gas. In the case of the tides, the horizontal scale is the radius of the planet \( a \) and the Doppler shifted frequency is \( u/a \), so that the vertical scale of the wave is \( u/N \). Putting this equal to \( h \) gives \( u = Nh \). He shows that for any forcing strength above a certain threshold value, the coupled system, mean flow plus tide, will settle on a mean wind speed near \( Nh \) and quite insensitive to the forcing strength.

The vertical momentum transfer in the tidal calculations by Newman and
Leovy is primarily between the level near 65 km, where the wave is driven, and approximately the 90 km level, where it is absorbed. The results nicely explain the dynamical structure of the top portion of the Venus atmosphere, at heights above about 50 km. The principal momentum balance is between vertical fluxes by the tides and by the mean flow. To an order of magnitude,

$$\overline{u'w'} \sim \overline{uw}.$$ (17)

Note that the tide and the mean Hadley circulation are both driven by solar heating, so the order of magnitude of $u'$ and $w$ are the same. Thus under this balance $\overline{u} = O(u')$. In reality Newman and Leovy find that numerical factors work in favor of the mean flow, and the mean flows that they calculate are several times larger than $u'$. Nevertheless this argument shows why vertical momentum fluxes by the thermally driven tides are unlikely to explain the mean flow at deep levels in the atmosphere. An estimate of the tidal velocity is given by Eq. (4) and is displayed in Fig. 10 (it is the same as the Hadley circulation estimate). At heights less than 50 km the velocities are a few m s$^{-1}$ or less.

Horizontal momentum fluxes by the tides are under a similar restriction. The balance Eq. (14) states the requirement for redistribution of momentum horizontally to maintain an angular momentum gradient that increases toward the equator, and we see that if $u' \sim \overline{w}$, it follows that $\overline{u}$ cannot greatly exceed $u'$.

The Newman and Leovy simulation evolves in time and is not restricted to periodic components. It produces an equatorward transport of momentum near the cloud base by free eddies, which are not restricted in amplitude in the same manner as the tides. They feel that the free eddies are probably not accurately simulated in this particular model, however, because the atmospheric stratification cannot be allowed to take on small values. The appearance of these eddies may be significant (see Sec. XV).

**XII. EDDY HORIZONTAL MOMENTUM TRANSFER MECHANISMS**

In a stratified atmosphere, eddies can transport momentum out of a mid-latitude jet and transfer it to lower latitudes. This has the effect of pumping angular momentum outward, toward the equator. Rossby (1947) speculated that "vorticity transport" would accelerate equatorial regions in a predominantly two-dimensional flow. One of the possible applications he had in mind was the Sun, which displays an equatorial angular velocity maximum. Gierasch (1975) speculated that this mechanism might be a key one on Venus, with the vertical transport by a Hadley cell being the other. The effect of eddy momentum transport in a spherical shell geometry was first quantitatively studied by Rossow and Williams (1979) in a numerical simulation. They found an inverse kinetic energy cascade, with a pumping of energy into two large-scale modes, rigid body rotation and a pair of eddies that formed a
wavenumber one disturbance. The phenomenon has since been documented in numerous three-dimensional computations (see, e.g., Del Genio et al. 1993; Suarez and Duffy 1992).

Although it may seem unphysical that eddies can transfer angular momentum equatorward against its gradient, it is not. Ordinary viscosity does this. The lowest energy state of a rotating disk of fluid, consistent with a given angular momentum, is rigid body rotation, with a maximum angular momentum density at the outer edge. The perplexing question is why momentum transfer occurs horizontally over large scales, rather than dissipating a mid-latitude jet locally, via either vertical eddies or small scale horizontal eddies. One key lies in the general properties of two dimensional flow. Fjörtoft (1953) proved that kinetic energy cannot cascade to small scales in two-dimensional flow, but must undergo an inverse cascade toward larger scales. Kraichnan (1967) described the physics of an inverse cascade in terms analogous to the well-known three-dimensional inertial cascade regime of Kolmogorov.

Numerical experiments by Herring and Mètait (1989) show that quasi-two-dimensional motions develop, with an inverse kinetic energy cascade, when the Richardson number based on the local fluid properties exceeds unity. This condition is met within the three major stable layers of the Venus atmosphere, which are centered near 15 km and 40 km, and within the stratosphere. The numerical experiments show horizontal flow confined to layers whose thickness is given by $u'\!/N$, where $u'$ is the fluctuation velocity. Within the Venus stable layers a typical $u' \approx 3$ m s$^{-1}$ and $N \approx 0.003$ s$^{-1}$, leading to a thickness estimate of about 1 km. This is consistent with the fluctuations displayed in Fig. 5a, although both the vertical and the velocity resolution of the Pioneer probe data is marginal for this conclusion. Possible origins of eddies on Venus are important questions. Baroclinic or barotropic instabilities are likely to exist near the mid latitude jet above the clouds (Young et al. 1984; Michelangeli et al. 1987). At deeper levels the mean structure is not well enough known to diagnose stability conditions. It is also possible that eddies are triggered by vertically propagating waves.

**XIII. BOUNDARY LAYER AND CONVECTIVELY DRIVEN WAVES**

Hou and Farrell (1987) propose that gravity waves are excited by convection within the lowest scale height of the Venus atmosphere, that the waves propagate vertically, and that those with horizontal phase velocity in the same direction as the mean flow are absorbed at critical levels, thereby accelerating the mean flow and contributing to its maintenance. One usually expects waves generated in the boundary layer to have small horizontal phase velocities and to contribute to drag rather than acceleration. This is the implication from the study of the convective region of the middle cloud deck by Leroy and Ingersoll (1995a), a region analogous to a surface boundary layer in the sense of being convectively active. Their study indicates that, as expected, convectively driven gravity waves typically have small phase speeds relative to the mean
zonal wind where they are generated. Because of vertical shear in the mean wind, waves carrying momentum that accelerates the mean zonal wind are absorbed within a few km of the convective zone, and therefore such waves cannot sustain the mean zonal wind beyond several km from the region of generation. Gravity waves which propagate obliquely to the zonal direction may have larger zonal phase speeds, but their contribution to the vertical flux of zonal momentum is small, and hence they also are not a significant source of momentum for the mean zonal wind. Nonlinear effects can considerably alter wave-mean flow interactions and change the location of critical levels (Fritts and Dunkerton 1984), but whether nonlinear processes would substantially enhance the role of convectively driven gravity waves in maintaining the superrotation is a subject for future study.

In the Earth's atmosphere there are examples of important exceptions to the general rule that boundary layer driven waves contribute more to wave drag than acceleration, as discussed above in Sec. IX. However, the wave phase speeds in the terrestrial case are comparable to the magnitude of the mean wind which they are supposed to drive (the quasi-biennial oscillation) because they have planetary-scale horizontal wavelengths. At tropical latitudes on Earth the effects of moisture can act to organize convection on a large scale and cause a coupling of convective activity to large scale waves (see, e.g., Holton 1992). Longitudinal variations in convective intensity may also contribute to forcing of long waves (Hitchman and Leovy 1988). On Venus it is not clear how large-scale organization might arise. Longitudinal variations of topography are a possible cause.

In addition to convection, boundary layer slope winds (Dobrovolskis 1993) or nonlinear eddies produced by the interaction of the mean flow with topography can potentially drive gravity waves into the upper atmosphere. Nonlinear calculations by Young et al. (1994) show that stationary waves induced near the surface can increase in amplitude with height, and may deposit momentum at higher levels. They also showed that nonlinear cascade of energy can generate dynamical structures with smaller scales than the originally excited waves. However, because the waves are stationary with respect to the planet surface, they act as a drag on the superrotation wherever they deposit momentum.

Consider a horizontal average of the angular momentum balance Eq. (14), in order to focus on the vertical momentum exchanges. The data suggests that the averaged rotation of the Venus atmosphere is predominantly, and perhaps exclusively, in one direction at all heights. If one makes the assumption that dissipative processes, such as wave drag, exist throughout the atmosphere, then the offsetting momentum flux from near-surface layers must be predominantly of one sense. In order to satisfy this requirement, forcing by westward waves would need to dominate over eastward if wave momentum transfer is to explain the superrotation. This is the requirement that the right-hand side of Eq. (14) be equal to zero on a time average. Thus, under the assumption that the average drag is oppositely directed to the average rotation, there must
be an anisotropy in the wave generation process if waves originating near the surface are to provide the forcing for the superrotation. This can come about, for example, because of feedback of waves or mean flow onto the convection field or because of special wave ducting properties of the atmosphere. None of these questions has been addressed in detail. To further examine the hypothesis by Hou and Farrell, the stratification, mean shear, and eddy properties in the lowest atmospheric scale height need to be better measured in order to determine wave ducting properties and the depth and intensity of convection.

**XIV. BOUNDARY LAYER FLOWS**

The solar flux reaching the surface is maximum at mid-day on the equator and is zero in the night hemisphere. Tomasko (1983), using Pioneer probe flux measurements and estimated optical properties of the atmosphere, presents an analytical approximate expression for the flux reaching the surface in the sunlit hemisphere as

\[ F \approx 80 \text{ W m}^{-2} (\cos \zeta)^{1/4} \]  

where \( \zeta \) is the solar zenith angle. The diurnal harmonic component on the equator is roughly 40 W m\(^{-2}\). The time dependent thermal forcing of the boundary layer can drive global tidal motions, global or local slope winds, and varying convection layer depth and strength that might in turn drive waves upward. The unknown infrared opacity and thermal structure of the deep atmosphere creates a major uncertainty in estimates of these boundary layer effects, because it is not known whether the mean profile is strongly stratified or not. If it is close to adiabatic, the diurnal thermal oscillation will be relatively deep and of small amplitude. If it is strongly stratified, the oscillation will be shallow and of larger amplitude.

The known amplitude of the forcing permits evaluation of the total buoyancy forcing, integrated through the diurnal boundary layer depth. In the case with heat advection unimportant, the heat equation reduces to

\[ \rho c_p \frac{\partial T}{\partial z} = -\frac{\partial F}{\partial z} \]  

where \( F \) is the upward heat flux. The heat storage in the surface is negligible (Gierasch and Goody 1970) and the forcing heat flux is the full surface insolation. For a harmonic forcing flux amplitude \( \Delta F \) with frequency \( \omega = 2\pi/(\text{solar day}) \approx 2\pi/(117 \text{ days}) \), the vertical integral of Eq. (19) gives

\[ \int T \, dz = \Delta T \Delta z = \frac{R T \Delta F}{c_p \omega p} \approx 1.2 \text{ K km} \]  

where \( p \) is the surface pressure and the boundary layer has been assumed to be thin enough so that the ambient density and pressure are approximately constant through it. The depth \( \Delta z \) is meant to be a measure of the boundary
layer thickness. Thus the forcing $\Delta F$, which is quite accurately known, fixes the product of the boundary layer depth and thermal oscillation amplitude, but neither the amplitude nor depth separately.

A lower limit on the boundary layer depth is given by the purely radiative case. For this limit to obtain, the boundary layer must be stratified with the Richardson number greater than $1/4$, so that turbulence is suppressed and all heat transfer is by radiation. The properties of a radiative boundary layer can be estimated for the case of a simple gray infrared transfer model. Hunten and Goody (1969) pointed out that the thermal optical depth in such a model can be estimated if the solar flux is known, by requiring that the greenhouse effect give the correct surface temperature. Using a solar flux at the surface of 20 W m$^{-2}$ we estimate an optical depth $\tau_R \approx 1600$. The opacity is $\kappa = \tau_R / H$, where $H$ is the scale height near the surface, and the heat Eq. (19) becomes

$$\frac{\partial T}{\partial t} = -\frac{\partial R}{\partial z} = \frac{\partial}{\partial z} \left( \frac{4\sigma T^3}{\kappa} \frac{\partial T}{\partial z} \right)$$

in the opaque limit (Goody and Yung 1989). Linearizing to examine the diurnal temperature perturbation for a thin boundary layer, one finds a radiative diffusivity

$$K_R = \frac{4\sigma T^3}{\rho c_p \kappa} \approx 0.02 \text{ m}^2 \text{ s}^{-1}. \quad (22)$$

Using $\kappa = \tau_R / H$ with an optical depth of 1600, a scale height of 15 km, a surface pressure and temperature of 90 bar and 730 K gives a diurnal boundary layer depth estimate of about 170 m, which implies from Eq. (20) an amplitude of about 7 deg. This estimate is probably a minimum depth and maximum amplitude, because most effects omitted from this model will act to increase the diffusivity. Nongray radiation would introduce smaller opacities over portions of the spectrum, and turbulence would enhance the heat transfer.

The boundary layer has potentially important effects on large-scale dynamics. It has been mentioned that turbulent convection in the boundary layer can be an important source of gravity waves. In addition, the semidiurnal boundary layer volume change is a forcing for an atmospheric tidal response that produces a quadrupole moment of the mass distribution and hence a net torque by the solar gravitational field (Dobrovolskis 1983, 1993; Dobrovolskis and Ingersoll 1980). The torque is in the correct sense to accelerate the observed atmospheric spin and was suggested by Gold and Soter (1971) as a possible cause of the superrotation. But the detailed calculation by Dobrovolskis and Ingersoll (1980) gives a net torque of about $1.8 \times 10^{16}$ J, corresponding to a surface stress of about $\tau_S \approx 2 \times 10^{-4}$ dyne cm$^{-2}$. As a rough estimate of a boundary layer wind necessary to create an offsetting stress, one may use the drag formula

$$\tau_S = C_D \rho U^2. \quad (23)$$
The density of the Venus atmosphere at the surface is about 0.06 g cm$^{-3}$. Using $C_D = 0.003$, appropriate for a neutral boundary layer a few meters above the surface (Priestley 1959) gives a flow speed estimate of $U \approx 1$ cm s$^{-1}$. Although this appears to be a small velocity, the tidal torque might possibly resolve an indeterminacy and define the sense of the mean flow, and it should not be ignored while our understanding is still incomplete.

Finally, Dobrovolskis (1993) shows that the diurnal boundary layer thermal disturbance will create up-slope and down-slope buoyancy forces that drive oscillating winds over topographic relief. Surface elevations and slopes are known from Pioneer Venus and Magellan radar data (Ford and Pettengill 1992). He calculates the winds for the case of a weakly stratified boundary layer, self-consistently estimates the turbulent diffusivity from mixing length theory, and then computes the associated boundary layer thickness and temperature amplitude. These come out to be approximately 5 km and 0.2 K. The surface stresses are on the order of 20 dyne cm$^{-2}$, which should be large enough to initiate saltation and soil movement (Iversen and White 1982). Observations, however, do not show correlation of aeolian features with surface slope direction (Greeley et al. 1994).

**XV. GENERAL CIRCULATION EXPERIMENTS**

Realistic numerical modeling of the Venus atmosphere is a challenging task. The range of time scales that must be described is very large. Dynamical activity, such as the tides, involves time scales less than a day. Thermal adjustment of the mean atmospheric structure involves time scales of decades. The range of length scales is also large. The tides and the mean flow are global in scale. Currently the minimum horizontal scale that must be resolved is not known because the key processes are not yet identified. In the vertical direction the resolution should be better than a scale height, and the circulation is more than eight scale heights deep.

Early three-dimensional numerical simulations of Venus-like atmosphere either did not produce rapid rotation of the atmosphere, or employed formulations of diffusion that rendered the results difficult to interpret (Källnay de Rivas 1975; Young and Pollack 1977; Rossow 1983). Recent work has focused on parametric studies based on terrestrial models. A few parameters or processes are altered, but the model is otherwise fixed in the terrestrial configuration. This approach avoids the problem of the extremely slow Venus thermal adjustment, and also permits easy comparison with the well documented and better-understood terrestrial case. Del Genio and Suozzo (1987) varied the rotation rate in a model driven only by a latitudinal heating gradient (no tides) and found that mid-latitude jets are prominent features of even slowly rotating regimes, but the model did not achieve strong equatorial superrotations. Del Genio et al. (1993) modified the radiative heating in the same model so that friction due to convective activity was reduced, and found that both equatorial and mid-latitude superrotation developed. The
mechanism is upward angular momentum transport by the Hadley circulation and easterly angular momentum transport by eddies generated by mid-latitude instabilities. In further work, De Genio and Zhou (1996) found that in terrestrial models having Venus' rotation rate the computed superrotation was sensitive to numerical precision because of the smallness of individual terms in the momentum balance. They showed that in general, the strength of superrotating winds increases with planetary rotation rate, but the efficiency of superrotation, i.e., the ratio of total angular momentum in the superrotating state to an atmosphere corotating with the solid planet, increases with decreasing planetary rotation rate.

Important results have also been obtained by workers who are primarily interested in the stability of the terrestrial circulation. Suarez and Duffy (1992) found that a very simple two layer numerical model can be induced into a superrotating state by imposing a longitudinally varying (but stationary) low latitude heating. There is a hysteresis effect, and the superrotating state can be maintained even when the amplitude of the triggering heating is reduced. Again the mechanism for superrotation is equatorial momentum transport by eddies and vertical transport by the Hadley circulation. These studies raise the interesting possibility that the Venus circulation pattern (and the Earth's) may not be steady and may have undergone historical variability, switching between a superrotating state and a "normal" state.

XVI. SUMMARY AND ASSESSMENT

The success story of the past two decades is the understanding that has been reached of the dynamics of the middle atmosphere region, from about 60 km to 90 km elevation. The PVO, in 1979, gave a first order characterization of the thermal and dynamical structure here. Fels and Lindzen (1974) had suggested that tidal momentum transports control the mean flow. Increasingly detailed theoretical work followed, by Peehman and Ingersoll (1984), Fels (1986), Leovy (1987), Baker and Leovy (1987) and Hou et al. (1990). Newman and Leovy (1992) calculated a balanced solution, showing tidal momentum fluxes balanced by mean flow advection. The predicted velocities and temperature perturbations are in good agreement with observations. We may conclude that to first order the dynamics of this height region is understood, at least at low and mid-latitudes where data is most complete. The high latitude polar vortex is not as well documented, and may contain surprises.

But the middle atmosphere dynamics turns out to have little to do with maintaining the overall spin of the Venus atmosphere. In fact, a major element of the tidal solutions is a downward eddy flux of momentum into the velocity maximum near 65 km elevation. This eddy flux acts to accelerate the cloud-top flow, but at the expense of higher regions, and is in fact in the opposite direction to the eddy fluxes that would be needed to pump momentum from the surface of the planet upward into the atmosphere. Fluxes of the latter type are also part of the tidal solutions, as pointed out by Fels and Lindzen, but
the detailed calculations have shown that tidal amplitudes below the clouds, where fluxes would be upward, are small. Tides cannot yet be ruled out as important beneath the clouds, but it seems unlikely.

The key to unlocking the mystery of the Venus atmospheric superrotation is in the momentum transports in the deep atmosphere. The crucial location is the region where the atmospheric spin increases from near corotation with the solid surface, within the lowest few km, to the height where a strong mean flow of about 50 m s\(^{-1}\) exists, at about 45 km elevation. In order to understand the flow, the balance between forcing and drag must be diagnosed and explained. Based on experience with the terrestrial atmosphere, wave drag probably operates on Venus and is likely to be the dominant drag to be overcome. To test this hypothesis, the nature of waves and turbulence needs to be determined. Two different classes of hypotheses for the acceleration mechanisms have been advanced. Under one, eddies or waves accomplish an upward momentum transport that drives the flow, and no other flow components are essential. Under the other, eddies or waves accomplish an equatorward transport and the Hadley circulation accomplishes the upward transport, and both components are essential. The two hypotheses are not mutually exclusive and a mix may be present.

Experience with geophysical fluid systems shows that theoretical modeling will not be conclusive except in conjunction with observations. We now know that the eddies or waves that are important in producing drag and in producing the momentum transports that offset it probably have amplitudes in the deep atmosphere between a few cm s\(^{-1}\) and a few m s\(^{-1}\) and are therefore beneath the accuracy threshold of the Pioneer Venus probe data. The thermal stratification is also crucial to establish, especially within the lowest 15 km where very little data exist. Specific goals for observation, based on the foregoing discussions, especially Sec. VII and Fig. 11, are:

1. **Stratification of the bottom scale height at several locations on the planet**, with an accuracy of at least 0.1 K and a measurement at least every 0.1 km in height. Processes constrained by this observation are boundary layer convection, waves generated by boundary layer convection, near-surface waves and wave propagation characteristics, and slope winds.

2. **Three velocity components in the bottom scale height at several locations on the planet** with an accuracy of at least 0.1 m s\(^{-1}\), profiled vertically with spatial resolution of 0.1 km. Two locations as near as possible to dawn and sunset are desirable to look for changes in convection intensity. Processes addressed are the same as for (1).

3. **Vertical scale (velocity coherence scale) and vertical structure (periodic or random) of eddies in the lower atmosphere**. Three velocity components are desirable, with accuracy of 0.1 m s\(^{-1}\), sampled at least every 1 km in height. Temperatures are desirable at the same spatial resolution and with accuracy of 1 K. Processes addressed are waves, turbulence, and momentum transfers. Gravity waves would be identified by partic-
ular phase correlations between velocities, and between velocities and temperatures, and also by spatially periodic structure.

4. Latitudinal and vertical structure of the zonal mean flow and the temperature. These are needed to determine the Richardson number and the horizontal shear, in order to define the stability properties of the mean flow. To determine the horizontal shear there should be at least three or four latitudinal points sampled in a hemisphere (north or south).

5. Relationship of eddy intensity to underlying topography or convection. Measurements (1)–(3) should be made at locations over both strong and weak topography. Processes addressed are wave drag, or wave generation in general, due to mean flow over topography or topographically induced slope winds.

6. Horizontal scale of eddies in the lower atmosphere. Measurements (2) and (3) should be made at different horizontal separations between about 200 km and 2000 km to determine the horizontal scale of dynamical activity.

For the purpose of understanding the superrotation, improved measurements of cloud microphysical properties and of radiative fluxes are not of high priority. A detailed understanding of the Venus greenhouse effect will require better information, especially on the thermal radiation field, but the heating imbalances that drive dynamics are quite well determined by the Pioneer Venus solar flux measurements and by the observed thermal field. Similarly, the Venus clouds present interesting and important questions of local cloud dynamics, microphysics, and chemistry, but these are probably not crucial issues for the superrotation.

Theoretical general circulation modeling, on the other hand, will be essential to assimilation of future data. Probe measurements can define the flow along only a few profiles, and only at certain instants of time. The statistical properties of the flow must be established by theoretical arguments that are constrained by these measurements. The tools will certainly include numerical general circulation computations. A hierarchy of these will probably be necessary, stressing different aspects of the flow. Boundary layer slope wind simulations, for example, may be necessary to isolate and treat with a single model, whose results are then abstracted and included in parameterized form in a global computation. Particular questions of importance that can only be answered by computations are the nature of the Hadley circulation, which is probably too weak at deep levels to be directly measured, and the global distribution of eddy momentum fluxes.

Acknowledgments. This work has been supported in part by the NASA Planetary Atmospheres Program.
REFERENCES


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