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COMPREHENSIVE SUMMARY OF AVAILABLE KNOWLEDGE OF THE METEOROLOGY OF MARS AND VENUS

by Edward M. Brooks

Prepared by
GCA CORPORATION
Bedford, Mass.
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ABSTRACT

This report presents a survey of available information on the meteorology of Mars and Venus. The emphasis is on the more recent observational and theoretical work, which has helped to clear up some of the uncertainties. Many uncertainties remain. Present indications of meteorological conditions on Mars and Venus are outlined below.

Mars

Carbon dioxide, which has been definitely identified spectroscopically in the atmosphere of Mars, is believed to constitute more than 50 percent of the Martian atmosphere. Formerly, this was believed to be true of nitrogen, which together with argon and/or neon is now believed to comprise the smaller remainder of the atmosphere, with the exception of other substances occupying .05 percent. There is probably no liquid water or liquid CO_2 on Mars, yet ice crystals of H_2O or perhaps of CO_2 may be responsible for the persistent blue haze, occasional white clouds, and the variable polar caps. Dust devils and dust storms, sometimes of hemispheric extent, are indicated by yellow clouds. Prevailing winds in summer are from an easterly direction and for the rest of the year are westerly, with a wave-type circulation pattern primarily in winter.

The atmospheric temperature, scale height, pressure, and density near the surface of Mars were all found from Mariner IV data to be lower than had been generally expected. This temperature has a mean value of roughly 200°K, but may vary more than 100°K diurnally and spatially between the equator and the winter pole. The scale height near the surface is about 9 km. The mesopause, at a height of about 100 km, has a temperature minimum which may be below 100°K, whereas the exosphere temperature could be above 500°K. At the surface, the mean atmospheric pressure and mass density average about 7 mb and 1.5×10^{-5} gm/cm^3, respectively. At all altitudes, the density is less than that of the Earth's atmosphere.

Mars appears reddish because its albedo decreases from 0.33 in the red to a minimum of 0.05 in the ultraviolet, probably due to powdered limonite in the light areas, which constitute 70 percent of the disk. The surface varies

about 6 km in altitude, but is relatively smooth on a large scale, and is densely covered with eroded meteoritic craters. The dark canals are believed to be tectonic ridges.

Venus

Venus is in very slow retrograde motion with a rotation period of about 243 days and the rotation axis within 10^{o} of the pole of its orbit. The climate is more diurnal than seasonal. Its 0.73 average albedo is very high, due to the reflection of its perpetual cloud cover, which looks almost always featureless except for rare markings visible at ultraviolet and near infrared wavelengths. Cusp caps and extensions of the horns of Venus are due to light scattered at the terminator by the cloudy atmosphere.

Venus has 4 gases which constitute all but .01 percent of its atmosphere. Nitrogen and neon comprise the great bulk of the atmosphere, CO_2 appears in small supply, and there is a trace of H_2O. The cloud layer is probably composed of H_2O ice particles of $7\frac{1}{2}$ to 10 microns in radius and some supercooled water drops near its base.

The general circulation of the atmosphere is theorized to be thermally direct: from the dark side to the light side of Venus at low levels, then upward, from the light side to the dark side at high levels, and finally downward on the dark side. The surface temperature, as revealed by microwave radiation, is very high: 610^{o} to $640^{o}K$ on the dark side and about $750^{o}K$ on the light side. At the top of the cloud layer, the temperatures determined from infrared measurements are about $240^{o}K$ on both the light and dark sides. Of the 3 classical (ionosphere, aeolosphere, and greenhouse) models of the atmosphere, the greenhouse effect of CO_2 and the cloud layer offers the best explanation of the microwave brightness temperatures. The pressure is of the order of 50 atmospheres at the surface, but only about 0.1 to 1 atmosphere at the top of the cloud layer.

The surface of Venus has a mean slope of 1/10 and is not so smooth as Mars, but smoother than the Moon. The dielectric constants indicate that the surface has no water, but some metals in the liquid phase may be present on the hot surface.

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1. INTRODUCTION

With the advent of the space age at the launching of the first artificial satellite in 1957, studies of the planets in the solar system were greatly stimulated along many lines of approach. Rapid strides are being made in both theory and observations over a wide range of the electromagnetic spectrum, but the literature presents many apparent inconsistencies, or contradictions. This summary will offer an appraisal of the conditions pertinent to the meteorology of Mars and Venus, as gleaned from papers selected from 935 references examined. The emphasis will be on the most recent works, which in some cases may supersede earlier research reports that have been compiled in other surveys, such as by Rea (1965) about Mars; Dickerman (1966) and Briggs and Mamikunian (1963) about Venus; or Van Tassel and Salisbury (1965) about both planets.

2. MARS

2.1 ATMOSPHERE

Although Mars has been telescopically observed near its oppositions for a couple of centuries, very little was learned about its atmosphere until the last two decades. Of particular importance were the recent spectroscopic observations and the observations of and by the Mariner IV satellite as it passed Mars at close range in July 1965, after the opposition of Mars in March 1965. The data lead to major revisions in Martian parameters, as described in many of the 1966 papers. Temperature, scale height, pressure, and density near the surface of Mars were all found to be lower than had been generally expected (Kliore, Cain, Levy, Eshleman, Fjeldbo, and Drake, 1966).

2.1.1 Composition

2.1.1.1 Gases and Vapors. — The gaseous components of the Martian atmosphere are generally detected and studied by means of their infrared emission and by their extinction of radiation. The latter includes their scattering and absorption of infrared radiation from the Martian surface, and of visible and ultraviolet radiation from the sun. However, in these regions of the spectrum, some gases such as nitrogen and the chemically inert gases are not optically active. To estimate the abundance of these gases, one can use the inferred astronomical history of Mars as an aid.

2.1.1.1.1 Changes in Atmospheric Composition: The composition of the atmosphere is probably different from what it was initially due to losses to space, losses and gains by chemical reactions at the surface, and gains by meteoric impacts and outgassing of the interior of Mars on a geological time scale. The losses of light gases from the exosphere would have been more severe for Mars than for the Earth because the values of Martian gravity and related molecular velocity required for gases to escape are relatively small. Any hydrogen generated by the ultraviolet photodissociation of water vapor would have escaped (Sagan, 1966) relatively quickly. In the photodissociation of CO_2, the generated atomic oxygen is also light enough to escape (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). In general, the rate of escape caused by thermal activity of molecules is insignificant due to the very low temperature of the Martian exosphere, but some gas may be eroded away due to bombardment by solar wind plasma (Fjeldbo, Fjeldbo, and Eshleman, 1966a). If a heavier primitive atmosphere of Mars had relative abundances of gases matching normal cosmic abundances, the present atmosphere should retain the heavier non-escaping gases, neon and nitrogen (Gross, McGovern, and Rasool, 1966).

An important atmospheric gain may result from a small asteroid striking Mars which would produce a volcano and would locally volatize the Martian crust. More CO_2 and water would be injected into the atmosphere than would be lost to space or through chemical reactions, provided the gas content of the crust was the same as that of a typical meteorite (Öpik, 1966).

2.1.1.1.2 Carbon Oxides: The two carbon oxides to be considered are CO_2 and CO. Carbon dioxide has been definitely identified spectroscopically in the atmosphere of Mars. In order to account for both the Mariner IV occultation observations and spectroscopic measurements, it is believed that CO_2 is the major constituent of the Martian atmosphere (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

The total amount of atmospheric CO_2 was determined by an investigation of the radiative absorption in weak unsaturated CO_2 lines. Thus, from data obtained during the 1965 opposition of Mars, abundance values of 65 ± 20 m-atm (Owen, 1966b) and 68 ± 26 m-atm (Belton and Hunten, 1966a) were found, the latter corresponding to a partial pressure of CO_2 equal to 5 ± 2 mb. Larger values were derived at an assumed temperature of $200^{\circ}K$: a CO_2 abundance of 90 ± 27 m-atm and a partial pressure of 6.6 mb (Spinrad, Schorn, Moore, Giver, and Smith, 1966).

The Mariner IV observations, however, indicated a total atmosphere of 13 to 16 gm/cm^2 (STP) (5-6 mb)(Öpik, 1966) which is close to the smaller amounts of CO_2 quoted above, hence suggesting that the atmosphere is largely

CO_2. From the observed refraction by the Martian atmosphere of radio signals from Mariner IV just before its immersion into occultation, the number density, mass density, and pressure of the atmosphere were found. Under the assumption of an atmosphere of pure CO_2, the parameters can be assigned to that gas, giving a CO_2 number density of $1.9 \pm 0.1 \times 10^{17}$ mol/cm^3, a CO_2 mass density of $1.43 \pm 0.10 \times 10^{-5}$ g/cm^3, and a CO_2 partial pressure of 4.9 ± 0.8 mb (Fjeldbo, Fjeldbo, and Eshleman, 1966a). The derived pressure at the emersion of Mariner IV was surprisingly high, especially for a pure CO_2 atmosphere: 8.4 ± 1.3 mb (Kliore, Cain, and Levy, 1966) (to be discussed in the section on pressure).

Another oxide of carbon in the Martian atmosphere is carbon monoxide. In the ionosphere, CO_2 can become positively charged as CO_2^+ or it can undergo photodissociation at altitudes of 70 to 80 km by solar ultraviolet radiation to CO and O (Shawhan, 1966), which can later recombine to CO_2 (Marmo, Shardanand, and Warneck, 1965). Carbon monoxide can also be produced by carbon dioxide joining with atomic oxygen: $O^+ + CO_2 \rightarrow O_2^+ + CO$. This reaction is more important than the reaction of atomic oxygen with either N_2 or O_2 because of the much greater abundance of CO_2 (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). The amount of CO produced from CO_2 by these two photochemical reactions may be only about 0.2 cm (STP) (Shimizu, 1966).

2.1.1.1.3 Nitrogen: Nitrogen was formerly believed to be the main constituent of the Martian atmosphere (Fjeldbo, Fjeldbo, and Eshleman, 1966a). It was hypothesized to make up the difference between the partial pressure of CO_2 and the total surface pressure (Rea, 1965) when the latter was thought to be much higher than the present estimates. The high atmospheric density derived from the Mariner IV occultation experiment would require, according to the equation of state, an incredibly low temperature if the composition were to be primarily nitrogen instead of CO_2 (Öpik, 1966), since N_2 has a molecular weight of only 28 as compared with 44 for CO_2. A low temperature of about - 100°C (Öpik, 1966) was derived from the Mariner IV occultation experiment on the assumption of a CO_2 atmosphere, but it was not as low as required for the density of a nitrogen atmosphere.

In the Martian ionosphere no N_2^+ emission was detected (Owen, 1966a). Yet cosmic abundance considerations suggest that the partial pressure of N_2 at the Martian surface should be about 1 mb (Gross, McGovern, and Rasool, 1966). This is also expressed by an atmospheric model having up to 20 percent of N_2 by volume (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

2.1.1.1.4 Nitrogen Oxides: There are many chemically possible oxides of nitrogen. Nitric oxide can be produced in the Martian ionosphere by the reaction, $O^+ + N_2 \rightarrow NO^+ + N$ (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966), but the predicted abundances of oxides of nitrogen are extremely low (Lippincott, Eck, Dayhoff, and Sagan, 1966). Also, no NO_2 absorption has been observed in Martian spectra, confirming previously set upper limits on its abundance (Owen, 1966b). This supports the implausibility of nitrogen oxides contributing to Martian phenomena (O'Leary, 1965).

2.1.1.1.5 Neon and Argon: Although these inert gases cannot be detected spectroscopically, cosmic abundances can be considered as a guide to estimates. Hence in the Martian atmosphere neon should exist and have a partial pressure of at least 2 mb at the surface (Gross, McGovern, and Rasool, 1966).

Argon is believed to be produced by the decay of radioactive potassium 40 in the crust (Loomis, 1965). This inert gas as well as nitrogen are surmised in the Martian atmosphere (Adelman, 1966) to account for the atmospheric density where the spectroscopically determined amount of CO_2 is insufficient to supply the total surface pressure (Rea, 1965). Argon has an atomic weight of 40, twice that of neon and nearly equal to the molecular weight of CO_2, which is 44. Various models of the atmosphere assume that argon can exist up to 20 percent by volume (Fjeldbo, Fjeldbo, and Eshleman, 1966a), up to 50 percent where the CO_2 abundance is 60 m-atm and the surface pressure is 10 mb (Gray, 1966a), and even up to 84 percent if the surface pressure is as high as 19.6 mb (Gray, 1966b).

2.1.1.1.6 Oxygen: Oxygen is important to Martian life if it exists. A simulated Martian environment with a low oxygen content was found to be sufficient to support some forms of plant and animal life (Hibben, 1965). The amount of ozone in an atmosphere is considered to be a sensitive indicator of the oxygen concentration (Paetzold, 1963).

It is quite likely that both atomic and molecular oxygen are produced at high altitudes in the Martian atmosphere. Atomic oxygen, along with CO, is a dissociative product of CO_2 and is distributed in the upper atmosphere by diffusion (Figure 1) (Fjeldbo, Fjeldbo, and Eshleman, 1966a). A rather definite upper limit of O abundance is 250 cm NPT (Marmo, Shardanand, and Warneck, 1965). Atomic oxygen becomes the principal atmospheric constituent above about 80 km because it has the lightest molecular weight (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). Theoretical ionization peaks due to O^+ occur near 95 and 150 km, in agreement with Mariner IV data (Shawhan, 1966). At 230 km, the density of O is about 100 times that of CO_2, but the atomic oxygen is converted rapidly to molecular oxygen by the reaction (Donahue, 1966), $O^+ + CO_2 \rightarrow O_2^+ + CO$. This is followed by a dissociative recombination of molecular oxygen ions with electrons, then the newly formed molecular oxygen may pick up a charge from an atomic oxygen ion (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966): $O^+ + O_2 \rightarrow O_2^+ + O$. Again, the molecular oxygen ion may be neutralized by an electron. Ultraviolet radiation can cause O_2 to photochemically dissociate to 2 O, which in the presence of a third molecule can recombine to O_2 (Marmo, Shardanand, and Warneck, 1965). The effect of diffusion should be considered in finding the distribution of the O_2 molecules as well as of the O atoms (Shimizu, 1966). The upper limit of O_2 abundance was set at 2 cm-atmospheres (Rea, 1965).

2.1.1.1.7 Ozone: Ozone is produced from atomic and molecular oxygen in the presence of M, a third molecule: $O + O_2 + M \rightarrow O_3 + M$; and it is destroyed by decomposition when radiated or by combination with atomic oxygen: $O_3 + h\nu \rightarrow O_2 + O$ and $O + O_3 \rightarrow 2O_2$ (Paetzold, 1963). The ozone

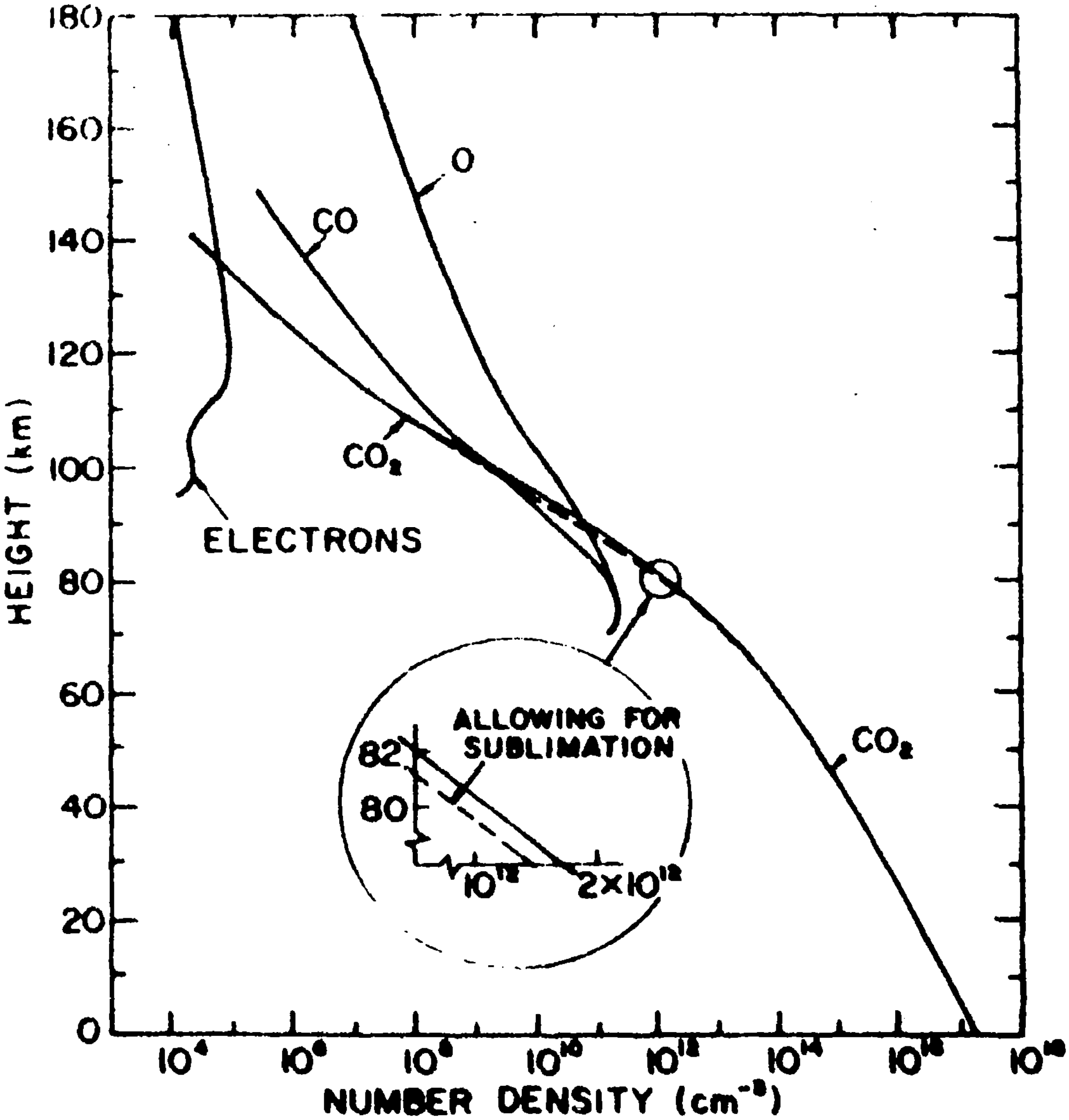

Figure 1. Number density versus altitude above Electris at the time of immersion into occultation. The limiting ion loss process in the main ionospheric layer is assumed to be $O^+ + CO_2 \rightarrow CO + O_2^+$ with a rate coefficient of 10^{-9} cm^3/sec.

distribution is derived from quantitative equations expressing these photochemical reactions. Since the amount of ozone depends more on the temperature than on the intensity of radiation, the heating effect of the absorption by the ozone itself must be considered (Paetzold, 1963).

There is no evidence for a theoretical ozone concentration peak at some altitude above the Martian surface, but only a monotonic decrease of O_3 concentration with increase of height (Marmo, Shardanand, and Warneck, 1965). Consequently, there is no temperature maximum aloft produced by ozone absorption of solar energy (Prabhakara and Hogan, 1965). More recently, the upper limit of the ozone abundance was reduced to 4 micron-atmospheres, corresponding to 2 cm-atm of O_2 (Rea, 1965).

2.1.1.1.8 Water Vapor: Water in its three phases is important to meteorology and also to life. This section is primarily concerned with H_2O as a vapor in the Martian atmosphere.

Spectroscopic observations show that Mars has only a trace of water, most of which may be on or under the surface, leaving for the atmosphere a small fraction as vapor amounting to 10-20 microns of precipitable water (Otterman and Bronner, 1966). Approximate values are 15 microns (Spinrad, 1966a) and 14 microns (Öpik, 1966). These measurements support the assumption that the average partial pressure of atmospheric water vapor is determined by the dissociation pressure of goethite ($Fe_2 O_3 \cdot H_2O$) on the surface (Adamcik, 1963). In addition to losses and gains of water vapor by chemical combination with and separation from goethite, water vapor subjected to ultraviolet radiation suffers losses by photodissociation into hydrogen and oxygen (Sagan, 1966).

Water vapor is not detectable in all Martian spectra, but only in spectra of certain regions with favorable local meteorological conditions (Münch, 1966). An important example is the spectroscopically determined water vapor migration away from a dwindling polar cap (Spinrad, 1966a).

2.1.1.1.9 Summary of Gaseous Composition: The gases and vapors can be grouped in three classes according to estimates of their abundances: (1) CO_2, A, Ne, N_2; (2) O, O_2, H_2O, CO; and (3) O_3, NO, NO_2, N_2O_4, and all others. The total abundances of the groups are of the order of magnitude of 100 m-atm, 5 cm-atm, and 5 micron-atm, or 99.95 percent, .05 percent, and .000005 percent, respectively of the total atmosphere. With considerable uncertainty, it can be stated that CO_2 accounts for about 70 percent, argon and/or neon about 20 percent and N_2 about 10 percent, but it is realized that CO_2 may constitute more than 90 percent of the atmosphere if the total pressure is found to approach the CO_2 partial pressure.

2.1.1.2 Atmospheric Suspensions. — Optical observations of the Martian atmosphere reveal airborne particles, which must be solid suspensions. Few, if any, liquids condense into drops because their saturation vapor pressures would exceed the low ambient vapor pressures of the same substances in the atmosphere. The solid particles must be small as shown by the long durations of their optical effects. They enter the atmosphere by wind transport off the surface, by encounter from space, and by sublimation from atmospheric vapors. Some particles eventually precipitate onto the surface while others evaporate within the atmosphere.

2.1.1.2.1 Optical Scattering by the Atmosphere: On the average, when no moisture is evaporating from the Martian surface, the atmosphere is transparent to long wave visible radiation (Miyamoto, 1966), and in general has a low value of optical thickness (Barabashov, 1965). In the short wave end of the visible spectrum, the atmosphere is more opaque, but it is still much too transparent (see table below) to be responsible for the hiding of surface features viewed through a blue filter (Dollfus and Focas, 1966b; Evans, 1965).

The measured illuminance of the atmosphere in sunlight exceeds that of Rayleigh scattering from a pure gaseous atmosphere having the same low surface pressure as that of the Martian atmosphere; the excess scattering is interpreted as being produced by a small amount of fine aerosols in permanent suspension (Dollfus, 1966a). Another indication of the presence of aerosols is that the fraction of the atmospheric extinction due to scattering is only about 20 percent, the remaining 80 percent being due to absorption (whereas in a pure atmosphere the scattering fraction would be typical of Rayleigh scattering by molecules, namely 100 percent of a much smaller extinction) according to Russian observations, as shown in the following table (Öpik, 1966):

TABLE 1
DISPOSITION OF SOLAR RADIATION IN THE MARTIAN ATMOSPHERE

Wavelength (Å)	Color	Transmission Coefficient	Scattering/Extinction
4600	blue	0.33	0.20
5200	green	0.54	0.22
5430	green-yellow	0.60	0.23
5800	yellow	0.69	0.24
6400	red	0.74	0.20

Because of small scattering fractions at all wavelengths, the amount of multiple scattering is not significant (Öpik, 1966).

2.1.1.2.2 Limb Darkening or Brightening: A planet with a uniform and isotropic surface albedo and no atmosphere should have a uniform brightness in sunlight, according to Lambert's law. If the planet has an atmosphere which absorbs the surface radiation and emits any kind of radiation, it will generally cause limb darkening or brightening (relative to the brightness nearer the center of the disk). This depends on whether the absorption or emission respectively, is the larger. In the case of thermal radiation, limb darkening or brightening is usually produced if the temperature decreases or increases, respectively, upward through the radiatively active layer. If the layer is sufficiently opaque, it may take the place of the surface in forming the apparent limb.

In the longer visible wavelengths, Mars appears limb darkened indicating an absorbing atmosphere (Öpik, 1966). In blue light, the limb darkening is not noted and may be replaced by slight limb brightening because the gain from the greater atmospheric scattering at the short wavelengths makes up for the loss by absorption. For example, bright marginal regions in blue-violet have been observed around a pole (Öpik, 1966). Trumpler corrected the derived Martian diameter from Wright's photographs in yellow to allow for possible limb darkening, and he also noted that limb darkening in the infrared was possible (Loomis, 1965). Some optical limb effects may be produced by anisotropic scattering when the albedo of a single scattering event is near unity (Morozhenko and Ianovitskiĭ, 1965). A series of photographic observations of a spot on the surface of Mars as it rotates yields a time cross section of brightness analogous to a space cross section, from which the degree of limb darkening or brightening can be measured. This method was used in photometric determinations of surface pressure on Mars.

2.1.1.2.3 Water and Ice: The Martian atmosphere contains only about 10^{-3} times the amount of H_2O found in the terrestrial atmosphere (Horowitz, 1966). Water substance is believed to be the ingredient of Martian white clouds (Wells, 1966). As the atmospheric water vapor pressures are less than the water-saturation vapor pressures, no liquid water clouds can exist in the Martian atmosphere, and the only atmospheric H_2O phase transitions are between vapor and ice (Otterman and Bronner, 1966). Moisture in the form of water vapor or perhaps ice crystals is transported discontinuously at successive mid-days in an equatorward moving wave of optical darkening of the surface maria (Otterman and Bronner, 1966).

White clouds have been identified as ice crystal mists (Dollfus, 1965), which often persist as veils in the atmosphere (Dollfus, 1966). To bring visual polarimetric and ultraviolet photometric estimates of surface pressure into agreement, Greenspan (1965) found the clouds must consist of 8×10^8 ice particles of 0.2 microns in diameter per square centimeter column (Cann, Davies, Greenspan, and Owen, 1965).

2.1.1.2.4 Dry Ice: Now that the amount of carbon dioxide vapor relative to the amount of water vapor is estimated to be higher than was formerly believed, the assessed probability for particles in the Martian atmosphere to be composed of solid CO_2 instead of H_2O ice has increased (Rea, 1965). Since some of the atmosphere is colder than the sublimation point of CO_2 (145°K at 4 mb), dry ice particles are expected to form (Fjeldbo, Fjeldbo, and Eshleman, 1966b). Some of the observed haze and particles may have been solid CO_2 particles (Eshleman, 1966) or possibly CO_2 particles may be an almost permanent feature at intermediate altitudes of the Martian atmosphere (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

2.1.1.2.5 Haze: Most haze and clouds are considered to be at altitudes below 30 km (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). Haze varies diurnally, tending to clear off by Martian noon and to reform in the late afternoon (Robinson, 1966). Haze also varies with locality and with season (Dollfus, 1965). There are two seasonal hazy situations: there is a white haze over the polar cap of the winter hemisphere, and there is a haze in middle latitudes during a period when the prevailing winds are shifting in late spring or early summer (Miyamoto, 1965).

A blue haze or violet layer is hypothesized to explain the strong atmospheric absorption (see table above) in the blue part of the visible spectrum (Dollfus, 1966b). The effects of blue haze are noticeable in the blue (Loomis, 1965) and prominent in the ultraviolet (Sagan, Phaneuf, and Ihnat, 1965). The absorption level of blue haze may be no higher than 10 km (Rakos, 1965), but there is some evidence for a violet haze at an altitude of about 100 km (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). According to Kuiper, the blue haze may be caused by sub-micron particles of CO_2, H_2O crystals, or dust acting as Rayleigh scatterers (Cann, Davies, Greenspan, and Owen, 1965; Rea, 1965). Blue haze has been explained by fine dust particles blown off the Martian surface by seasonal winds (Sagan and Pollack, 1966a) or arriving as interplanetary dust from space, as suggested for terrestrial noctilucent clouds (Palm and Basu, 1965). Another theory of the blue haze is that it is produced by solar protons ionizing the Martian atmosphere, but other theories are recognized as possible (Blum, 1965).

2.1.1.2.6 Blue Clearings: The clearing of blue haze is usually explained by the removal of particles producing it. With reference to the last three theories of blue haze presented above, blue clearings could be caused by the settling or deposit of fine dust particles back on the surface (Sagan and Pollack, 1966a); by a diminished influx of interplanetary dust, suggested by a small negative correlation between blue clearings and meteor showers (Palm and Basu, 1965); or by deflection of the solar wind away from Mars by the Earth's magnetic field (Blum, 1965). The last theory was a consequence of the observed high frequency of blue clearings near the opposition of Mars. Since the Earth's field is not strong enough to cause the alleged deflection of solar protons, another mechanism might be a turbulent hydromagnetic shock wave from the Earth which might cause sufficient instability in the Martian wind to clear away the particle layer (Blum, 1965). A simpler hypothesis to account for the occurrence of blue clearings at Martian opposition does not require any reduction in particle density, but nearly an optical effect at the phase angle of 0°, whereby the Martian surface features become brighter and offer improved optical contrasts to the viewer. This hypothesis was offered independently by Menzel and Öpik (Menzel, 1963).

2.1.1.2.7 Clouds: A relatively light area on the limb of Mars in a photograph was interpreted as a cloud with its top at 25 km (Leighton, 1966), below the 30 km altitude specified above for haze and clouds (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). The brightening of clouds at the limb is attributed to frost and mist in the Martian evening(Miyamoto,1965).

Cloud observations furnish useful information about surface relief and wind patterns (Miyamoto, 1966; Loomis, 1965). Map plots of cloud positions have topographical implications, such as a W-shaped cloud system over mountain ridges aligned in a W pattern (Wells, 1966).

During the Martian spring there is a W-E cloud system in middle latitudes analogous to the terrestrial polar front structure (Miyamoto, 1965; Ohring, Tang, and Mariano, 1965). In summer there are cloud patterns extending from the polar cap to beyond the equator, but in fall all the cloud patterns disappear (Miyamoto, 1965).

Observations have been made of Martian clouds of at least 4 colors: blue, white, gray, and yellow, which will be discussed in that order. Rea (1965) defined types of clouds according to these colors, but omitting gray.

Although blue haze is common, blue clouds are rare. Blue and white clouds are believed to consist of water ice or CO_2 ice particles (Rea, 1965).

Some white clouds are of orographic origin (Rea, 1965). In general, they exhibit two peaks of maximum occurrence during a Martian year, the white cloud histogram corresponding to a map of seasonal water vapor transfer between the poles across the equatorial regions (Wells, 1966). According to the presumed location of moisture, white clouds may occur over the equator in a belt or over the winter pole, the white clouds being more common over the N than S polar cap (Miyamoto, 1965). Selected occurrences of white clouds may reveal the possible presence of volcanoes (Wells, 1966).

A gray cloud of unusual color spread 1500 km to the Martian limb after an "explosion" on January 16, 1950 (Öpik, 1966). This is an example of one of 30 cases of gray cloud, which have been attributed to volcanic activity on Mars (Katterfeld, 1965). However, the thinness of the atmosphere speaks against past eruptions, and hence arguments that the maria are covered by wind-driven volcanic ash are not convincing (Öpik, 1966).

Yellow clouds exist on different scales, some occupying local areas within light desert areas, others covering entire deserts (Wells, 1966). During a Martian year, they seem to have three peaks of occurrence, the principal one being generally several weeks after the perihelion of Mars (Wells, 1966), when the lower atmosphere is warmest and most unstable. Sometimes sporadic events of extended yellow clouds or haze last for days, weeks or even longer (Öpik, 1966). A good example of this was the great yellow cloud of September, 1956 (Miyamoto, 1966). The observed frequency of the yellow-veil phenomena could be explained by a major meteoric collision, which theoretically should occur with an average frequency on the order of 6 years (Öpik, 1966). The yellow color of the clouds spectroscopically matches the color of limonite, which is believed to be present on the Martian surface (Loomis, 1965). This suggests that the clouds are composed of dust, the subject of the next section.

2.1.1.2.8 Dust: Dust clouds, with their tops normally extending to the tropopause, consist of solid particles with diameters of more than about 20 microns (Anderson, 1966). They appear as yellow clouds (Dollfus, 1965; Loomis, 1965) which were windswept up from bright areas (Rea, 1965). Such clouds tend to form at low latitudes in the winter hemisphere (Ohring, Tang, and Mariano, 1965). Veils of fine powders often persist after a dust storm (Dollfus, 1966b). The finest particles seem to linger in the atmosphere over the entire surface, as for example in the 1956 major dust storm, which left increased haziness and decreased optical contrasts for months afterwards (Öpik, 1966). According to photometric and polarimetric observations, the extra scattering produced by dust particles indicates that there are about 10^{15} aerosol particles with radii of the order of 10^{-2} microns in the atmosphere above a surface area of 1 sq. cm (Ianovitskiĭ, 1965). The lower wavelength limit of visibility of surface details varies with the dust count between λ 4260 A and λ 4580 A.

Another source of dust besides the Martian surface is interplanetary space (Palm and Basu, 1965). The presence of a meteoric dust layer in the Martian atmosphere was indicated by the apparent enlargement of Mars in 1926 (Wells, 1965a).

2.1.2 General Circulation

Low values of surface pressure and water vapor pressure eliminate on Mars most weather problems of the Earth except for those related to the winds (Tepper, 1965). Information on the circulation of the Martian atmosphere is quite incomplete. The horizontal motion is best revealed by tracers, such as aggregates of suspensions, provided the individual particles are small enough to have a negligible terminal velocities of fall relative to the atmosphere. However, the tracers do not portray motions at elevations above and below their level. This is where theories of vertical profiles of the wind velocity vector can come in handy. At best, only large scale circulation patterns can be observed, but motions on smaller scales might be inferred theoretically.

2.1.2.1 Wind Speeds and Directions.—

2.1.2.1.1 Motions of Clouds and Dust: In the summer hemisphere, clouds develop locally along the dark fringe surrounding the polar cap, drift in prevailing easterly winds to low latitudes along NE-SW lines, cross the equator, turn toward the E and move with winter hemisphere westerly winds (similar to terrestrial anti-trade winds aloft), then finally dissipate (Miyamoto, 1965).

Clouds in the Prepontis region of Mars, where they are common, indicate upward motion of thermal convection, caused by solar heating of the surface, Prepontis being the hottest place on Mars according to radiometric measurements (Miyamoto, 1965). As a small Prepontis cloud grows into a massive cloud layer, it is observed to drift with summer easterly winds, modified by the thermally direct local circulation (Miyamoto, 1965). White clouds, statistically common over desert regions bordering dark regions, may remain stationary for some time, suggesting that some of them may be lee-wave clouds (Wells, 1965b),which indicate a wind direction from the mountains or highlands causing them. However, their identity as lee-wave clouds is considered rather dubious (Öpik, 1966).

During the history of Mars, repeated windblown dust storms must have occurred to account for the observed aeolian erosion of craters (Anders and Arnold, 1965) until they were eventually obliterated by a light colored dust layer (Loomis, 1965). Dust as well as water vapor may account for the wave of darkening, which by this hypothesis is caused by a wind of sufficient strength and duration to remove the dust cover, leaving the darker surface underneath visible (Wells, 1966). Wind erosion destroys most tectonic lines on Mars, but develops some of those located along courses of vapor migration into canals (Miyamoto, 1966).

Windspeeds required for raising dust are not as strong on Earth as on Mars (Anderson, 1966), where they must be greater than 145 km/hour and where actual windspeeds may even exceed 300 km/hour (Rea, 1965). Such high winds can only be of short duration, such as in relatively small disturbances (Ryan, 1964). The presence of dust clouds suggests that there are tornadoes (Öpik, 1966) or dust devels greater than 100 meters in diameter, which require the temperature lapse rate to be sufficiently great for thermal convection to occur (Neubauer, 1966). Observations of cloud motions show that prevailing windspeeds of 60 km/hour are common (Loomis, 1965) and that they sometimes reach close to 100 km/hour (Rea, 1965; Gifford, 1964).

2.1.2.1.2 Surface Winds: Theoretical values of maximum surface windspeeds on Mars may exceed 100 m/sec (360 km/hr) and of average meridional wind components are larger than those of the Earth (Tang, 1965). There is a seasonal variation of windspeeds associated with the annual variation of the poleward temperature gradient. In summer, prevailing winds are too weak to lift dust off the light colored low surfaces, but they can remove dust from the dark highlands, whereas the prevailing winter winds are strong enough to lift dust off both kinds of surfaces (Pollack, Greenberg, and Sagan, 1966).

Prevailing surface wind directions at middle latitudes are westerlies during most of the year, but become easterlies during the Martian summer (Miyamoto, 1966). Between the spring westerly and summer easterly wind regimes there is a transitional period of weak and rather chaotic air currents (Miyamoto, 1965). Meridional wind components transport moisture from the waning polar cap in spring to lower latitudes (Otterman and Bronner, 1966). The meridional speeds of isopleths of water vapor concentration are in good agreement with the rate at which the wave of darkening spreads equatorward (Ohring, Tang, House, and Mariano, 1966). The rate of travel, consistent with surface windspeeds, is for a few hours daily, totaling 45 km per day according to deVaucouleurs (Otterman and Bronner, 1966). In some well defined channels, the rate of equatorward progression of darkening is less than half this speed (Loomis, 1965). The evidence for a true movement rather a nearly simultaneous outbreak of darkening is not compelling (Pollack, Greenberg, and Sagan, 1966).

2.1.2.2 Fields of Motion. —

2.1.2.2.1 Circulation Patterns: Circulation on a planetary scale is proven to exist by the hemispheric spreading of the clouds from the January, 1950 explosion and of the September,1956 dust storm (Öpik, 1966). An interaction between the circulations of the N and S hemispheres is indicated by the transport of water vapor from one pole to the other (Loomis, 1965). A diffusion model for interhemispheric water vapor exchange requires a large scale eddy diffusion coefficient (Ohring, Tang, House, and Mariano, 1966).

Wind patterns are expected to change markedly with season (Sagan and Pollack, 1966a). As an axially symmetric zonal circulation around a pole cannot remain stable, in the mean for a Martian year it must be replaced by a wave-type circulation (Tang, 1966). The seasonal mean conditions can be represented by symmetric westerly flows in spring and fall, and wave circulations from the W in winter and E in summer (Miyamoto, 1963). The mean zonal motion at an altitude where the atmosphere pressure is 1/4 that at sea level theoretically is about 36 m/sec (Tang, 1966). Theoretical estimates of mean large scale zonal and meridional windspeeds and of large scale vertical velocities are larger than the corresponding terrestrial values (Tang, 1965).

If a triangular dark area represents ash deposits from a volcano at a vertex, the observed curvature of certain dark areas indicates a trajectory of the tracers in a curved flow (McLaughlin, 1956) characteristic of cyclones or anticyclones. Areas which appear bright blue, with less than normal optical absorption due to a scarcity of aerosols, indicate subsiding polar anticyclones (Öpik, 1966), but this is contrary to experience on Earth, where anticyclones generally have excess pollution. Martian cloud observations suggest the existence of subtropical anticyclones, meridional flow at upper levels and frontal phenomena at equatorial latitudes (Tang, 1965). The presence of a springtime polar front as a feature of the N hemisphere circulation (Miyamoto, 1966) is indicative of the horizontal convergence and lifting of air masses originating in different anticyclones.

2.1.2.2.2 Vortices and Convection: Tornadoes are believed to be responsible for lifting large volumes of dust (Öpik, 1966). In dust devils, Martian windspeeds are theoretically about 80 percent of terrestrial speeds (Neubauer, 1966). The steep average temperature lapse rate and the weak gravity on Mars are more favorable for initiating dust devils there than on the Earth (Neubauer, 1966). Enormous upward convection accompanied by strong local winds near Martian mid-day (Weil, 1965) seems more likely than small convective effects near the surface (Kachur, 1965). The fact that the frequency of yellow clouds increases following increases in insolation and windspeeds points to a stronger convection (Wells, 1966).

2.1.3 Temperature Distribution

The surface of Mars has a thermally rigorous climate characterized by near-cryogenic temperatures at night and large, seasonally dependent temperature variations during the day (Kachur, 1966a). Because of the extreme spatial and time variability of temperature from spectral intensity measurements, a single temperature value becomes more useful when the position of the object measured and the Martian season and time of day of the measurement are specified. The wavelength or frequency of the measurement is also pertinent information because often electromagnetic energy in different parts of the spectrum penetrates to different depths in the planet's atmosphere or surface.

2.1.3.1 Temperature. —

2.1.3.1.1. Surface Temperature: Passive microwave observations of the Martian surface were made, by which an average disk temperature at 3.4 mm was found to be $190 \pm 40^{o}K$ (Epstein, 1966); other measurements yielded 240 $(+72, -48)^{o}K$ at 94 Gc (3.2 mm) and 230 $(\pm 42)^{o}K$ at 35 Gc (8.6 mm)(Tolbert, 1966); also a value of $225 \pm 10^{o}K$ at 8 mm was obtained (Kutuza, Losovskii, and Salomonovich, 1966). At a longer wavelength of 21.2 cm, the highest average effective brightness temperature was reported, namely $271 \pm 76^{o}K$ (Davies and Williams, 1966). These temperature values all seem too high compared with Kellermann's summary of 1965 measurements (Kellermann, 1965): $192 \pm 26^{o}K$ at 6 cm, $162 \pm 18^{o}K$ at 11 cm, and $190 \pm 41^{o}K$, which are all less than $200^{o}K$. The last results show that disk integrated thermometric determinations by microwave observations gave significantly lower temperatures than corresponding infrared surface temperatures, because the radio emission originated below the Martian surface (Sagan and Pollack, 1966b).

Theoretically highlands are only a few degrees cooler than low bright areas because of the small slopes of the Martian surface and the small greenhouse effect of the atmosphere (Sagan and Pollack, 1966f).

Latitude effects on temperature are quite large. When Mars was at aphelion, the temperature at the limb in the equatorial regions was determined to be $254^{o}K$, as contrasted with $201^{o}K$ at the pole; the subsolar point (local noon) had a temperature of $273^{o}K$; and the average temperature, weighted according to the areas of latitude zones, was $250^{o}K$ (Pettit, 1961; Sinton, 1961). These observations do not support the theory of a symmetric zonal flow, which requires a smaller poleward temperature gradient (Mintz, 1961a). The edge of the darkening wave is supposedly located at the latitude where the diurnal maximum temperature is $0^{o}C$, namely at latitude 22^{o} at the winter solstice according to Mintz (Otterman and Bronner, 1966).

The occultation of Mariner IV by Mars provided an opportunity to measure the atmospheric refractivity and density, from which the temperature could be found. The results showed that the Martian atmosphere was considerably colder than was previously anticipated (Fjeldbo, Eshlemann, Kliore, Cain, Levy, and Drake, 1966). At the point of immersion of Mariner IV, at latitude 55° (Öpik, 1966) in the S hemisphere, where it was late winter, the derived temperature was -98°C (175°K); at the point of emersion in local late summer, the temperature was -37°C (236°K) (Adelman, 1966). For an atmosphere consisting of all CO_2, the immersion temperature was computed to be 180±20°K, the other value of 175±25°K being for an atmosphere of 20 percent argon and 80 percent CO_2 (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

2.1.3.1.2 Seasonal and Diurnal Temperature Changes: The latitudinal temperature gradient on Mars is small near the time of the summer solstice (Pollack, Greenberg, and Sagan, 1966) due to the solar heating of the summer pole. Between latitude 50° and the pole, the seasonal decrease of the diurnal minimum temperature brings it by wintertime to below 145°K, the sublimation temperature of CO_2 at a partial pressure of 4 mb, and at the pole itself, to below 100°K, thereby making the formation of dry ice possible (Leighton and Murray, 1966). Another type of seasonal effect arises from the relatively large ellipticity of the orbit of Mars. The changing distance of Mars from the sun causes a considerable variation of mid-day temperatures from 259-269°K at aphelion to 295-305°K at perihelion (Kachur, 1966a).

Sunrise temperatures, observed at the terminator, are in the vicinity of 170-210°K (Kachur, 1966a). At the equator, the surface temperature may drop to -50°C (223°K) even in the summer (Strughold, 1966). A day on Mars near the time of perihelion may be characterized by temperatures such as the following: a diurnal minimum of -60°C at dawn, -49°C at 0700, 17°C at 1000, and a maximum of 39°C at 1400, according to Sinton and Strong (Öpik, 1966). The corresponding daily average temperature is -22°C, but would be -28°C for Mars at its mean distance from the sun, and the diurnal temperature range is 101 C° (Öpik, 1966), enormous by terrestrial standards.

Daytime temperatures of the dark maria are about 8 C° higher than the temperatures of the light deserts due to the greater absorption of insolation by the maria (Öpik, 1966). The diurnal temperature variations between night and day are smaller in moisture patches, which increase the thermal inertia of the soil (Otterman and Bronner, 1966). Diurnal temperature ranges vary between 100 and 150°C, the latter corresponding to a soil of poor conductivity (Leighton and Murray, 1966).

2.1.3.1.3 Scale Height $\left(\frac{R^*T}{mg}\right)$. The scale height is directly proportional to the absolute temperature and inversely proportional to the product of the molecular weight and the acceleration of gravity. Its variation with height in the lower atmosphere of Mars depends almost entirely on the temperature lapse rate, while in the upper atmosphere it is affected also by changes in the molecular weight associated with changes in atmospheric composition with increasing height.

Because of its low temperatures, the Martian atmosphere is confined to near the surface (Fjeldbo, Fjeldbo, and Eshleman, 1966a), which can be represented by a small scale height. The low temperature together with the large molecular weight (44 for CO_2) compensate for the 62 percent lower gravity than that of the Earth in the scale-height determination (Fjeldbo, Fjeldbo, and Eshleman, 1966a). A scale height of 9.3 km corresponds to a mean temperature of 203°K for a pure CO_2 Martian atmosphere (Belton and Hunten, 1966a). The Mariner IV occultation experiment yielded a small surface scale height of 9±1 km (Fjeldbo, Fjeldbo, and Eshleman, 1966a) and a low altitude of the main ionospheric layer (Fjeldbo, Eshleman, Kliore, Cain, Levy, and Drake, 1966). The scale height at the emersion point of Mariner IV was about 11.5 km, which was larger than the 9 km value at immersion (Kliore, Cain, and Levy, 1966). There were no obvious changes of scale height up to at least 30 km, according to the occulation experiment results (Sloan, 1966). In the Martian ionosphere, the electron scale height above the electron peak at an altitude of 125 km was about 25 km, indicating an unexpected low temperature (Sloan, 1966). In one model of the ionosphere, the scale height at an altitude above an ionization peak at 140 km was only 17.2 km, which was less than the 20 to 25 km determined from Mariner IV; and in another model of the ionosphere, the scale height was 23 km at a point above a major ionization peak of 150 km (Shawhan, 1966). All of these results emphasize the coldness of the Martian atmosphere.

2.1.3.2 Thermal Stratification. —

2.1.3.2.1 Troposphere: Measurements of the Martian atmosphere by means of infrared spectroscopy in the CO_2 band at 1.05 microns gave a rotational temperature of 194±32°K, in agreement with the mean temperature of 203°K from the scale height (Belton and Hunten, 1966a). The effective altitude was 12 km, below which the lapse rate was assumed to be 4.5 K°/km (nearly equal to the adiabatic rate, 4.7 K°/km) down to the surface, where the temperature was about 55 K° higher (Belton and Hunten, 1966a). Other determinations of the 1.05 micron band gave 200 ± 50°K with an adiabatic lapse rate below (Owen, 1966a); and the lower values of 160 to 180°K, which are remarkably close to the Mariner immersion value of 180°K (Hunten, 1966). Moreover, very low temperatures between 30 and 100 km are required to provide for the observed low densities at the height of the ionosphere above (Fjeldbo, Fjeldbo, and Eshleman, 1966a). A set of model atmospheres, most of which had a surface temperature of 230°K and an adiabatic lapse rate, yielded temperatures between 140°K and 160°K at altitudes near 50 km (Prabhakara and Hogan, 1965; Ohring and Mariano, 1966).

The vertical changes of temperature inferred from the Mariner IV occultation experiment were a small lapse rate up to 20 km at the immersion point, typical of the high latitude in winter (Mintz, 1961) and a greater lapse rate at the emersion point, where it was summer (Fjeldbo, Fjeldbo, and Eshleman, 1966a). For summer dust storms, the surface air must be unstable (Neubauer, 1966), with a lapse rate greater than the adiabatic lapse rate. In a model of the lower atmosphere, convective equilibrium (an adiabatic lapse rate) was assumed to exist up to the tropopause (Anderson, 1965).

2.1.3.2.2 Tropopause: The height of the tropopause varies inversely with the surface temperature. For a surface temperature between 230° and 270°K, the tropopause would be located below 10 km (Prabhakara and Hogan, 1965). A 210°K surface temperature is consistent with a tropopause at 14 km and at a temperature of 140°K (Johnson, 1965). The height of a dynamically produced tropopause was computed to be at about 20 km (Tang, 1965). The height of the Martian tropopause can be placed at approximately twice the height of the Earth's tropopause (Ohring, Tang, and Mariano, 1965), which would place it some 15 to 30 km above the Martian surface. The tropopause usually marks the upper limit of convection (Anderson, 1966), but it is poorly defined because of the small difference between the average lapse rates above and below it.

2.1.3.2.3 Stratosphere: Unlike the terrestrial stratosphere, which usually has a temperature inversion just above the tropopause and a temperature maximum at a greater altitude, the Martian stratosphere has neither one of these properties. There is a steady decrease of temperature with height and there is not enough ozone or carbon dioxide to produce a temperature maximum by the absorption of ultraviolet or infrared radiation, respectively (Prabhakara and Hogan, 1965). Above the tropopause the average lapse rate is about 0.64 C°/km (Johnson, 1965).

2.1.3.2.4 Mesopause: There is some disagreement as to how low the mesopause temperature minimum is. It has been assigned the value of 155°K (Prabhakara and Hogan, 1965). Alternatively, the temperature minimum has been placed below the sublimation point of CO_2 (Fjeldbo, Fjeldbo, and Eshleman, 1966b), which is less than 145°K, since the mesopause partial pressure of CO_2 is less than 4 mb. The height of the mesopause is about 100 km, where the temperature minimum may be as low as 85°K (Johnson, 1965) or even 50°K (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

2.1.3.2.5 Thermosphere: It is doubtful whether the thermosphere exists as a distinct layer in the Martian atmosphere. One opinion is that an F_2 ionosphere is so cold that there can be no increase of temperature above the mesopause, the lapse rate being assumed isothermal (Johnson, 1965). Another opinion is that the thermosphere is a layer extending from 100 to 140 km, characterized by a temperature inversion, with an increase of temperature from 50°K at the base to 80°K at the top of the thermosphere (Figure 2) (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

2.1.3.2.6 Ionosphere: The base of the Martian ionosphere is at an elevation of about 100 km (Fjeldbo, Fjeldbo, and Eshleman, 1966a), and consequently the ionosphere overlaps the thermosphere if the latter exists. For 3 models, each with one main ionospheric layer analogous to a terrestrial E, F_1, or F_2 layer, the temperature is on the order of 400°, 200°, or 100°K, respectively (Fjeldbo, Fjeldbo, and Eshleman, 1966b). In the height range from 120 to 200 km, the temperature is less than 200°K (Kliore, Cain, Levy, Eshleman, Fjeldbo, and Drake, 1965). The lapse rate may be isothermal at a temperature of 85°K (Adelman, 1966) or isothermal above the 140 km level

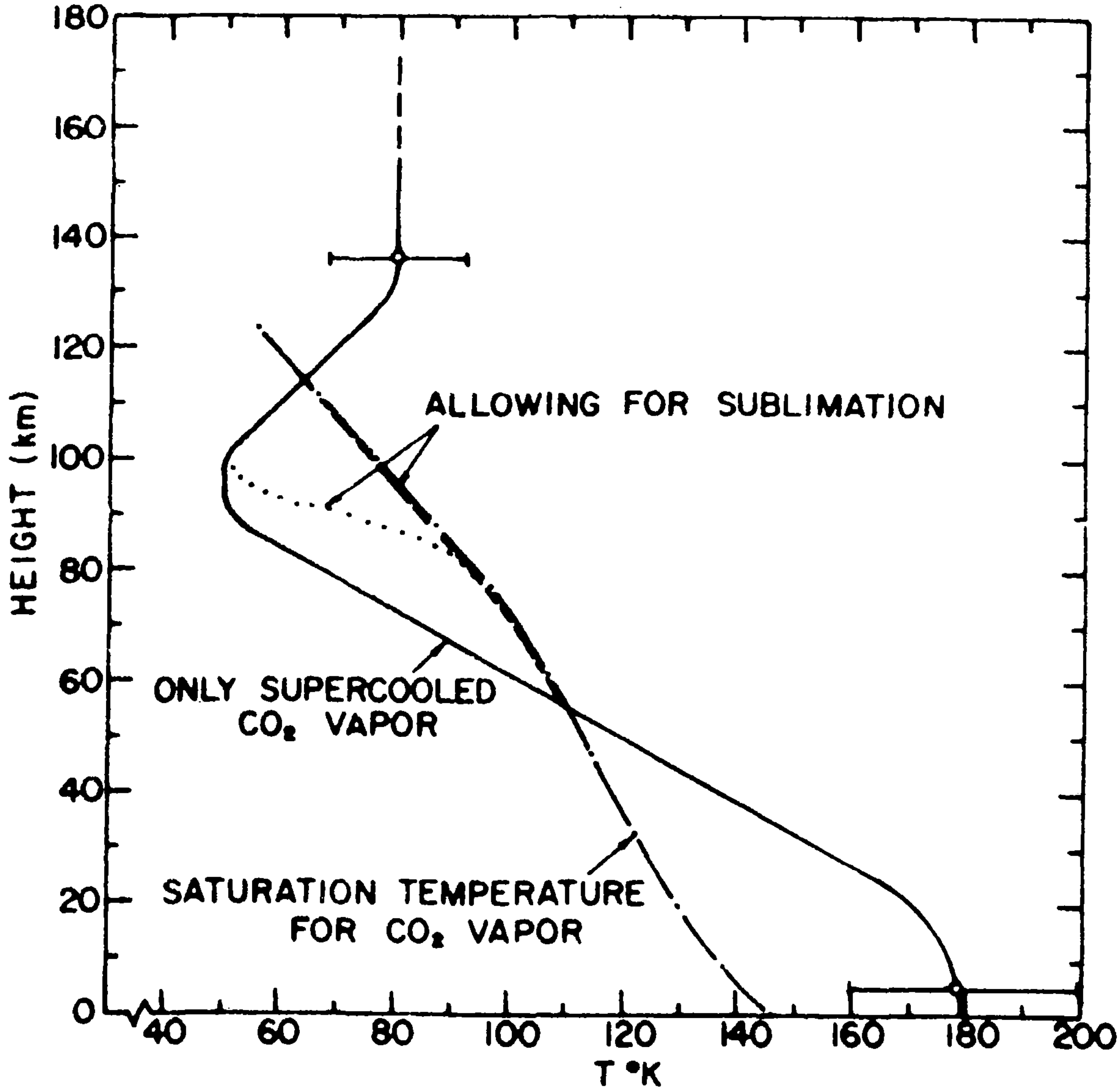

Figure 2. Temperature versus altitude above Electris at the time of immersion into occultation. The limiting ion loss process in the main ionospheric layer is assumed to be $O^+ + CO_2 \rightarrow CO + O_2^+$ with a rate coefficient of 10^{-9} cm^3/sec.

at 80^{o}K, based on the electron scale height (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

2.1.3.2.7 Exosphere: A wide range of temperature values have been considered as appropriate for the exosphere, which extends upward from a height of about 140 km (Fjeldbo, Fjeldbo, and Eshleman, 1966a). One model assigns it a temperature of 400^{o}K (Donahue, 1966). Based on a pure CO_2 atmosphere and a surface pressure of 8 mb, a temperature with a large uncertainty was calculated: 550 ± 150^{o}K (Gross, McGovern, and Rasool, 1966). Another possible exospheric temperature range is 600 to 1200^{o}K, controlled in part by the carbon monoxide abundance (Sagan, 1965).

2.1.3.3 Heat Budget.—The Martian atmosphere absorbs and scatters radiation by means of its gaseous constituents and solid particles. From radiation charts prepared to show the net energy flux through a horizontal surface of a model atmosphere, it was learned that self absorption within a gas limits significantly the net amount of radiation it emits (Myer, Ohrenberger, and Thompson, 1965). The random Elsasser band model used for defining the absorption by the strong 2-micron bands of CO_2 gave a satisfactory account of the observed atmospheric transmission (Gray, 1966b).

Mars has an absorbing atmosphere, indicated by limb darkening, particularly in the violet, where 85 percent of the extinction is due to absorption and 15 percent due to scattering by dust (Öpik, 1966), contrary to the less credible view that the existence of an absorbing layer on Mars in scarcely possible (Sharanov, 1965). Due to the low atmospheric pressure, however, there is only a slight greenhouse effect (Sagan and Pollack, 1966f).

If radiative equilibrium calls for a large lapse of temperature with increase of height, vertical convection destroys the equilibrium (Belton and Hunten, 1966a) and reduces the lapse rate to the adiabatic rate. This conversion of a radiative equilibrium to a convective equilibrium usually is confined to the lower atmosphere, the upper atmosphere remaining in radiative equilibrium. The resulting vertical temperature profile, which is the basis for a convective-radiative equilibrium model, is not very sensitive to variations in surface pressure or carbon dioxide concentration (Ohring and Mariano, 1966). By reasoning this conclusion in the reverse order, one can see that the meteorology of Mars, even at high altitudes, depends significantly on the surface temperature (Anderson, 1965).

Mars has a unique microclimate at the surface, unlike that of the Earth and Moon (Kachur, 1965). Due to the tenuous atmosphere, the heat fluxes at the surface are essentially radiative (Öpik, 1966). Of the total infrared radiation lost in 24 hours, 72 percent is emitted during the daytime and 28 percent at night (Öpik, 1966) owing to the big difference between day and night surface temperatures. The microclimate depends on how fast the surface temperature changes not only from the solar and Martian radiations, but also due to the thermal inertia of the soil (Kachur, 1966b). A layer of micron-sized dust particles on the surface is expected to have a lower thermal conductivity than the CO_2 atmosphere because the dust particle size is smaller than the mean free path of CO_2 molecules (Kachur, 1965). About 90 percent of the direct solar radiation incident on the Martian atmosphere is absorbed by the surface, and the loss of infrared radiation to space is 85 percent of the blackbody flux corresponding to the surface temperature (Kachur, 1965). The result is a net gain of radiative heat, which is lost by conduction into the atmosphere and the soil, the heat conduction amounting to less than 10^5 ergs cm^{-2} sec^{-1} (Kachur, 1965). Some radiogenic heat may originate in the interior of Mars (Weertman, 1966).

In the high atmosphere, heat is carried into the relatively cold layer between 30 and 100 km in height by conduction from the warmer lower atmosphere and thermosphere; meanwhile, additional heat is obtained from the absorption of ultraviolet radiation and X-rays, and heat is lost by the radiative emission of atmospheric CO, O, and CO_2 (Fjeldbo, Fjeldbo, and Eshleman, 1966a). Carbon monoxide is not so important a thermostat as was formerly believed (McElroy, l'Ecuyer, and Chamberlain, 1965). Some solar protons, not deflected in the absence of a Martian magnetic field, enter the atmosphere and their energy is converted to kinetic energy of neutral molecules (McElroy, l'Ecuyer, and Chamberlain, 1965). A comparison of the ultraviolet and visible solar energy absorption by O_2 and O_3 with the infrared absorption by CO_2 shows that they are of comparable magnitude below 30 km, but the CO_2 absorption is considerably larger than the O_2 and O_3 absorption above 30 km (Prabhakara and Hogan, 1965). A model atmosphere which fits the Mariner IV results has a surface temperature of 200°K, a lower atmosphere up to 80 km in radiative equilibrium, and the upper atmosphere of CO_2 dissociated into CO and O, assumed to be in diffusive equilibrium (Gross, McGovern, and Rasool, 1966).

Apparently, there will still be many details that need to be worked out for the heat budget of Mars when new data are available.

2.1.4 Pressure and Density

The mass of the Martian atmosphere and its distribution are important parameters, which can be represented quantitatively by pressure and density values. However, they cannot be measured as directly or precisely as the temperature, revealed by the intensity of the radiation in selected wavelength intervals.

2.1.4.1 Pressure.—Until about 1963 the surface pressure on Mars was assumed to be about 80 mb based on photometric and polarimetric observations (Owen, 1966b). The most satisfactory photometric observations were made of the brightness versus wavelength of selected surface spots at changing viewing angles, as Mars rotated, to determine how much effect the atmosphere had as the optical path length changed (Hunten, 1966). The amount of light with a wavelength variation or with a polarization typical of Rayleigh scattering was extracted from the results to derive the atmospheric pressure (Hunten, 1966). Sytinskaya and also de Vaucouleurs derived a surface pressure of 90 mb, even when some allowance was made for scattering from the permanent haze (Hunten, 1966). Dollfus, by making refined measurements of polarization, obtained 85 mb, but subsequently corrected the surface pressure to 40 mb by allowing more precisely for the reflection from the surface (Hunten, 1966).

Spectroscopic determinations of surface pressure are more reliable, since the results are not sensitive to scattering by solid particles. The spectroscopic technique involves the measurement of atmospheric absorption in a weak unsaturated CO_2 band, which reveals the CO_2 abundance, and the absorption in a pressure-broadened strong saturated CO_2 band, which gives the product of the CO_2 abundance and the total pressure. Dividing the latter by the former yields the pressure. For a given pressure-abundance product, the derived pressure is inversely proportional to the CO_2 abundance. The failure to detect a weak CO_2 band can be used to infer a corresponding maximum possible CO_2 abundance and consequently a minimum pressure from the observed absorption by a strong CO_2 band, according to Goody (Hunten, 1966). The weak CO_2 band at 8689Å, detected by Sinton, could not have been observed if the CO_2 abundance were as low as that corresponding to a pressure as high as 85 mb (Hunten, 1966). From the absorption of this band, Kaplan, Münch, and Spinrad (1964) found a CO_2 abundance of 55 m-atm, which was higher than previously believed, and a lower surface pressure of 25 ± 15 mb. From the same spectrogram Moroz (1964) obtained an even lower pressure value of 15 mb.

It is apparent that the spectroscopic measurements give pressures considerably less than the 80 mb determined by photometric and polarimetric observations, but there is still a large uncertainty in the spectroscopic values, based on 3 very weak CO_2 lines (Owen, 1966b). Kuiper (1964) explained the big disagreement between the old and new values as due to the scattering particles in the Martian atmosphere. If there were no aerosols, a pure atmosphere with a pressure between 80 and 90 mb would give the same photometric intensities as actually observed (Dollfus, 1966b). Even the polarimetric determination of 30 mb is too high because of scattering by pulverized limonite on the Martian surface (Dollfus and Focas, 1966a).

The most recent spectroscopic observations give lower pressure values than the older spectroscopic values. Spinrad, Schorn, Moore, Giver, and Smith (1966), using a CO_2 abundance of 90 ± 27 m-atm obtained surface pressures of 8 ± 4 mb and 16 ± 8 mb from the 1.6-micron and 2-micron strong CO_2 bands, respectively, the former being considered more reliable than the latter by Kuiper, the observer. From a CO_2 abundance of 60 to 85 m-atm, Gray (1966b) derived a surface pressure of 7.1 ± 2.2 mb. Spencer (1966) chose 6 mb as his preference for surface pressure, based on the 8689Å data of Spinrad and Schorn and on the 1.6-micron data of Kuiper and Owen, This 6-mb value is also preferred by Belton and Hunten (1966a), who found that a total pressure of 5 to 6.7 mb is consistent with a CO_2 partial pressure of 5±2 mb corresponding to an abundance of 68±26 m-atm. They established the minimum surface pressure as equal to the lowest CO_2 partial pressure of 3 mb, i.e., 5 - 2 mb (Belton and Hunten, 1966a).

Another method of obtaining surface pressure was by means of the measurements of Mariner IV just before and after it was occulted by Mars. The observed refraction of the radio waves from Mariner IV gave the atmosphere's vertical gradient of density, to which the speed of the waves is proportional, above the apparent points of immersion and emersion of Mariner IV. Kliore, Cain, and Levy (1966) reported surface pressures for a pure CO_2 atmosphere and for an atmosphere containing 80 percent CO_2 and 20 percent argon as follows: 4.9 ± 0.8 mb and 5.2 ± 0.8 mb, respectively, for the immersion point, and 8.4 ± 1.3 mb and 8.8 ± 1.3 mb, respectively for the emersion point. The large difference (exceeding 50 percent of the immersion point pressure) between the immersion and emersion pressures was the most interesting result of the Mariner IV occultation experiment (Eshleman, 1966). The measured Martian radii at immersion and emersion were 3384 ± 3 km and 3379 ± 4 km, respectively (Kliore, Cain, and Levy, 1966). The higher pressure and smaller radius at the emersion point indicates a lower gravitation level there than at the immersion point (Eshleman, 1966). Because there is no visible reference level for Mars, such as sea level for the Earth, the main uncertainty is the height of the immersion or emersion level with respect to the poorly defined mean level, according to Fjeldbo, Fjeldbo, and Eshleman (1966a). Even this mean level may not be an appropriate reference level for pressure because the solid surface appears to be an ellipse flattened about twice as much as the equipotential surface (Loomis, 1965). A graph of the logarithm of the pressure up to a height of 180 km above the surface (Figure 3) shows that the greatest fractional decrease of pressure per unit height was at a height of about 90 km, where the temperature was lowest, namely at the mesopause (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

The Mariner IV surface pressure results seemed too low compared to some of the spectroscopically determined values (Münch, 1966), but were in agreement with the latest spectroscopic values indicating 6 mb. There were some problems in the interpretation of the Mariner IV observations: (1) the use of a scale height constant with respect to altitude, (2) Fresnel diffraction of radio waves at the limb of Mars, (3) uncertainty in the height of the terrain, and (4) the possibility of significant ionization down to the surface (Münch, 1966). A surface temperature of 230° instead of 180° would increase the derived surface pressures by 25 percent which is equivalent to the effect of decreasing the elevation by 5000 ft, according to Münch (1966). Because of these problems, his choice of surface pressure is 8 to 10 mb, based mostly on spectroscopic observations. The difference between the infrared spectrometrically determined pressure of about 12 mb and the pressure from Mariner IV of about 6 mb may nevertheless be mutually consistent, because the infrared value refers to an average over the bright and dark surface areas whereas the Mariner IV is biased toward the high elevations of the immersion and emersion points (as indicated by their proximity to dark areas) (Sagan and Pollack, 1966f). If the Mariner IV value of 6 mb is typical of the dark highlands, the surface pressure of the bright lowlands should be about 20 mb to give an average which will agree with the infrared value of 12 mb (Sagan and Pollack, 1966f).

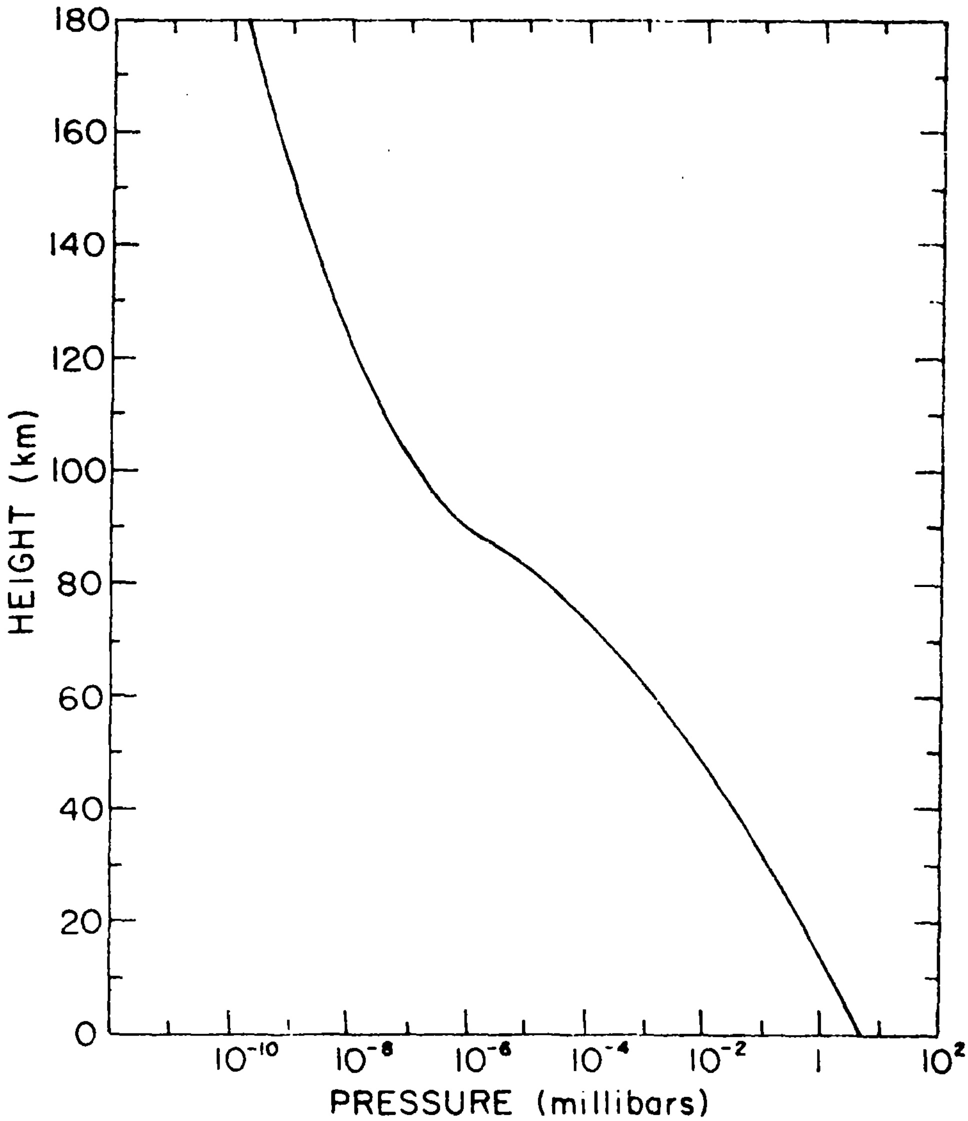

Figure 3. Pressure versus altitude above Electris at the time of immersion into occultation. The limiting ion loss process in the main ionospheric layer is assumed to be $O^+ + CO_2 \rightarrow CO + O_2^+$ with a rate coefficient of 10^{-9} cm^3/sec.

If solid CO_2 is deposited every half Martian year on alternating polar caps, there would be a Martian semiannual variation in the partial pressure of CO_2 vapor, with the lowest values occurring in the winter of each hemisphere, when part of the CO_2 vapor has been removed from the atmosphere to the polar cap in the form of dry ice (Leighton and Murray, 1966). This periodic fluctuation is computed to amount to ± 1 mb either side of the average CO_2 partial pressure at the surface, and should also appear as a semiannual fluctuation of the same magnitude in the total surface pressure, since it is the sum of the partial pressures of the gaseous constituents at the surface (Leighton and Murray, 1966). The hypothesized Martian semiannual variation of the surface pressure is somewhat asymmetrical about a mean of 3.7 mb as graphed (Figure 4) by Leighton and Murray (1966). It shows a principal minimum during the S hemisphere winter because of the larger amount of dry ice deposited with the lower temperature at the higher altitude of the S pole (as compared with the N pole in winter), and it shows a principal maximum 3 Martian months later, which is one month after the perihelion of Mars, when the dry ice becomes completely sublimated into the atmosphere due to excess solar heat received by Mars.

A comprehensive paper on the problem of the pressure in the Martian atmosphere up to 1965 has been prepared by Cann, Davies, Greenspan, and Owen (1965).

2.1.4.2 Density.—The Martian atmosphere causes refraction of electromagnetic waves. Its refractivity is approximately 3.7 "N" units (Sloan, 1966). The density of the atmosphere can be derived from its refractivity since they are directly proportional. The most useful refractivity measurements were made from the radio waves of Mariner IV just before and after its occultation by Mars (Fjeldbo, Fjeldbo, and Eshleman, 1966a), and additional measurements were made photoelectrically of the light from the Martian moon, Phobos, in connection with its eclipses by Mars (Rakos, 1965).

The density of the Martian atmosphere was found to be even lower then previously estimated. Early in its astronomical history, Mars may have had a dense atmosphere (Anders and Arnold, 1965). The present density of the Martian atmosphere at the surface was obtained over Electris from Mariner IV before its immersion into occultation (Kliore, Cain, Levy, Eshleman, Fjeldbo, and Drake, 1966):

TABLE 2

SURFACE DENSITIES IN 2 MODELS OF THE MARTIAN ATMOSPHERE

ATMOSPHERIC COMPOSITION	100% CO_2	80% CO_2 + 20% Argon
Number Density	$1.9 \pm 0.1 \times 10^{17}$ mol/cm^3	$2.1 \pm 0.1 \times 10^{17}$ mol/cm^3
Mass Density	$1.43 \pm 0.1 \times 10^{-5}$ gm/cm^3	$1.55 \pm 0.1 \times 10^{-5}$ gm/cm^3

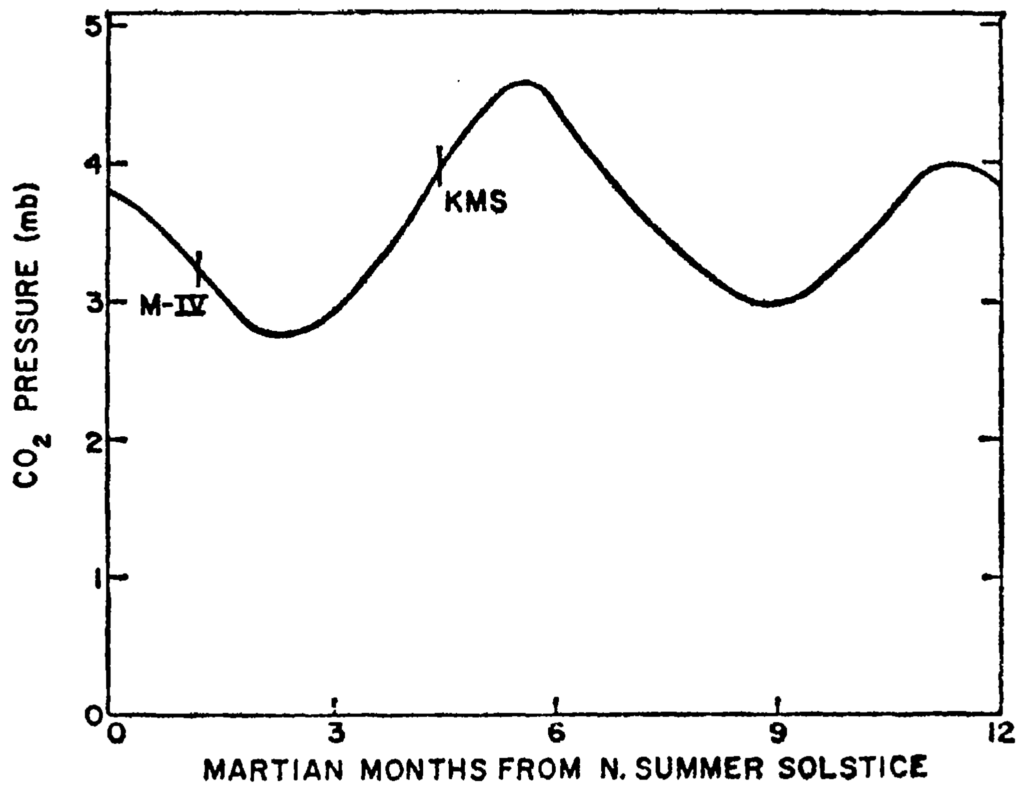

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Figure 4. Variation in CO_2 partial pressure, due to seasonal condensation and evaporation. The Martian date for the measurement of CO_2 pressure and total pressure by Kaplan, Münch, and Spinrad (KMS) (12) and of total pressure by Mariner IV (M-IV) (18) are indicated.

In the tenuous CO_2 Martian atmosphere, the molecular number density at the surface is only about 0.8 percent of the Earth's surface value and equal to the value at a height of about 34 km above the Earth's surface (Fjeldbo, Fjeldbo, and Eshleman, 1966a). The surface mass density for Mars is about 1.2 percent of that for the Earth, a higher percentage because the average atmospheric molecular weight is greater for Mars than for the Earth (Kliore, Cain, Levy, Eshleman, Fjeldbo, and Drake, 1965). Mariner IV density determinations were subject to the same uncertainties as the pressure determinations. The principal uncertainty was the height of the occulting surface feature relative to the mean height of the Martian surface (Fjeldbo, Fjeldbo, and Eshleman, 1966a). A less important source of density error was the presence in the atmosphere of dust particles, the density of which was estimated to be 3 gm/cm^3 (Anderson, 1966). Ionization near the surface could also cause an erroneous increase of number density by not more than 0.3 percent (Fjeldbo, Fjeldbo, and Eshleman, 1966a).

The density of the Martian atmosphere is assumed to decrease exponentially with an increase of height (Sloan, 1966). The number density diminishes by 1 percent for every 90 meters of height increase (Fjeldbo, Fjeldbo, and Eshleman, 1966a). The density aloft in a model atmosphere depends on the choice of surface temperature except at the height of 10 km, where the mass density is 5.77×10^{-6} gm/cm^3, irrespective of the assumed surface temperature (Anderson, 1965). The atmospheric density for Mars was found to be less than that for the Earth at all altitudes (Figure 5) (Fjeldbo, Fjeldbo, and Eshleman, 1966a) in spite of the small acceleration of gravity on Mars. The density of the Martian atmosphere decreases by nearly 10 orders of magnitude from the surface to the base of the exosphere (Fjeldbo, Fjeldbo, and Eshleman, 1966a), derived from the low values of refractivity found in the outer atmosphere from the Mariner IV radio waves. At ionospheric heights, the number density can be found from the electron density profile (Fjeldbo, Fjeldbo, and Eshleman, 1966a). For 3 models of the ionosphere, each with a layer analogous to the E, F_1, or F_2 layer of the Earth, the corresponding mass densities are in the ratios of 10^4, 10^2, and 1 (Fjeldbo, Fjeldbo, and Eshleman, 1966b). A model with a F_2 layer seems to fit theory and observations best (Fjeldbo, Fjeldbo, and Eshleman, 1966c). No ionosphere was detected from Mariner IV after its emersion from the dark limb, where night was occurring (Sloan, 1966).

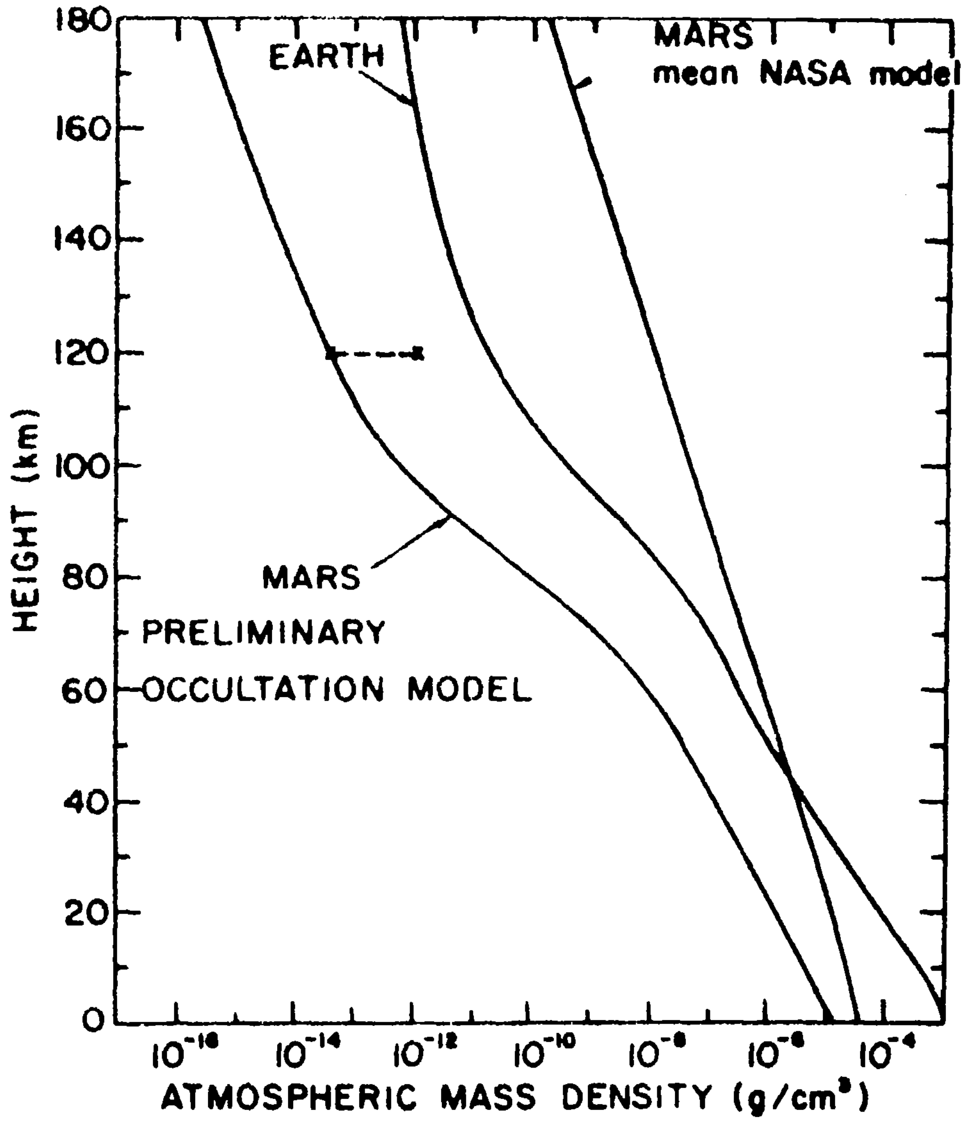

Figure 5. Atmospheric mass density versus altitude above Electris at the time of immersion into occultation. The NASA engineering model shown here was used for planning purposes prior to the Mariner 4 mission.

2.2 SURFACE

The surface of Mars is important to Martian meteorology because of its many interactions with the atmosphere. With wide fluctuations in temperature, the surface exerts a controlling influence on the temperature of the lower atmosphere by radiation and conduction. Highland areas localize thermal convection and orographic lifting, thereby causing an adiabatic cooling of the Martian air and sometimes the formation of clouds. Martian winds lift solid particles off the surface to produce dust storms, and they lose momentum by frictional drag over rough terrain. Evaporation and condensation of H_2O and CO_2 on the surface have noticeable effects, particularly in the polar regions.

2.2.1 Appearance of Mars

The interpretation of optical or radio observations of planets is usually ambiguous. In the case of Mars, the scarcity of optical material and its lack of confirmation are especially disturbing (Öpik, 1966). Nevertheless, attempts must be made to find meaning in pictures, with the realization that a hypothesis may have to be modified or replaced when new observations are available or when new theories give better answers to old questions.

2.2.1.1 Albedo. —

2.2.1.1.1 Visible Spectrum: The albedo of a surface area is related to the temperature since it equals unity minus the absorptivity, the transmissivity being zero. The surface is about 3/4 covered by bright orange colored areas, the remaining 1/4 being dark areas, which are subject to seasonal changes in color tone (Loomis, 1965). In more precise terms, 70% of the surface constitutes the orange, yellow, or reddish "continentes"; the dark maria, 27%, and the remaining 3% are transitions areas (Öpik, 1966). The bright areas are concentrated in the N hemisphere, the dark areas in the S hemisphere, and seasonal white areas in the polar regions (Rea, 1965). The dark areas in the S hemisphere are largely within 30^{o} of latitude from the Martian equator (Loomis, 1965). The Mariner IV pictures showed extensive spottiness in surface detail, consisting of small light and dark elements with linear dimensions of 10-100 km (Öpik, 1966). One simple model for dark areas is an intermixture of blue-green material of low reflectivity with normal bright area materials, each forming about 1/2 of the surface area (Loomis, 1965). But maria may contain more dark elements, possibly to near saturation, whereas continentes may contain a mixture of bright and dark elements in different proportions, causing a greater uneveness in surface brightness than is characteristic of maria (Öpik, 1966). The dark areas are sometimes described as greenish or bluish, but are actually of a brownish grey shade which is less reddish than the continentes (Öpik, 1966). In the maria small dark objects seem to be added to a general orange background (Öpik, 1966). Temporary color centers may be produced by electromagnetic and corpuscular solar radiation impinging on certain solids of the Martian surface (Smoluchowski, 1965).

The red color of Mars is due to a decrease of the albedo toward shorter wavelengths of the visible spectrum, as illustrated by the following values (Öpik, 1966): 0.33 at λ 6900 A (extreme red), 0.18 at λ 5500A (green-yellow), and 0.046 at λ 3600 A (near ultraviolet). A graph of reflectivity versus wavelength (Fig. 6) (Rea, 1965) shows that at this last wavelength the albedo is a minimum, then increases up to 0.08 at shorter wavelengths in the ultraviolet. If allowance is made for atmospheric absorption and scattering, the externally measured albedo can be modified to represent the reflectivity of the Martian surface itself, which turns out to be larger than previously quoted albedo values at all wavelengths (Öpik, 1966): 0.53 at λ 6400 A (red), 0.40 at λ 5800 A (yellow), 0.34 at λ 5430A(green-yellow), 0.29 at λ 5200Å(green), and 0.25 at λ 4600 A (blue). As viewed outside the Martian atmosphere, the geometrical albedo of bright areas rises rapidly with increasing wavelength throughout the visible spectrum, the same being true of dark areas to a lesser extent (Loomis, 1965).

There was no decisive difference between the spectra of continentes and maria as measured by Mariner IV (Öpik, 1966). The darkish areas appeared reddish like light areas (Rea, 1965). At wavelengths longer than λ 5500 A (yellowish-green) the albedo of bright areas is about 2 times that of dark areas, the maria being relatively very dark in the near infrared; but at shorter wavelengths the ratio becomes significantly less than 2 in the green and nearly 1 in the blue (Loomis, 1965). The small overall albedo of Mars contributes to a low optical contrast of the surface (Goy, 1966). The lack of contrast is particularly notable in photographs taken in blue light (Rea, 1965). Surface markings are visible through green, yellow, and red filters, their contrasts increasing at longer wavelengths (Öpik, 1966). The disappearance of markings at short wavelengths is due to atmospheric dust (Öpik, 1966) rather than an intrinsic decrease of surface reflectivity with decreasing wavelength (Evans, 1965), because the wavelength variation of contrast is too great for the latter interpretation. The view that the red color of Mars is mainly due to its atmosphere (Kozyrev, 1964) appears untenable because of the weak absorption of the atmosphere.

2.2.1.1.2 Radar: Early radar observations of the Martian disk showed that its radar cross section varies from 2% to more than 10% of the geometric cross section, which indicates that the surface of Mars is significantly flatter than that of Venus (Goldstein and Gillmore, 1963). From radar relectivity measurements at 12.5 cm wavelength along the parallel of 21°N Martian latitude (Goldstein, 1966a),mapping of the surface roughness showed that Mars in certain areas is quite smooth, but some rough, strongly reflection regions have also been found (Goldstein, 1965a). Mars on the whole appears smoother than Venus (Goldstein, 1966a)and also other planets, Mars having a marked degree of surface differentiation (Pettengill, 1965). Recent measurements of radar cross section of Mars averaged 8.6% (Goldstein, 1966a),and ranged from 3 to 13% of the geometrical cross section, the higher values being observed when dark regions of Mars were in the region of radar reflection (Dyce, 1965). In observations during both the 1963 and 1965 oppositions of Mars, a very striking association was noted between large radar reflectivity and the presence of optically dark areas (Sagan, Pollack, and Goldstein, 1966).

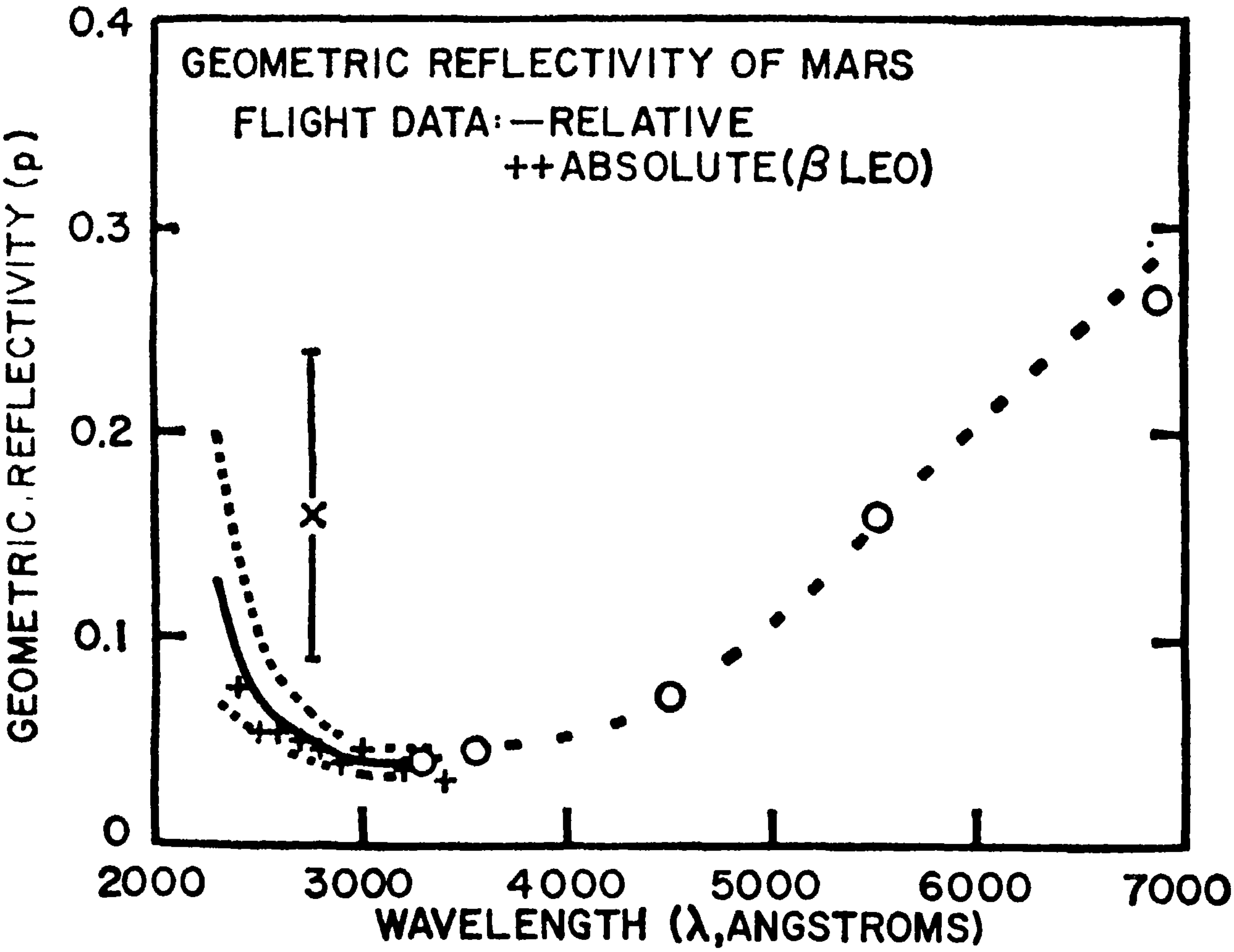

Figure 6. The geometric reflectivity of Mars. Solid line, relative data adjusted to 0.04 at 3400Å; +, absolute reflectivity determined by comparison with β Leo, plotted independently. The dashed lines below 3400Å represent the error range applied to the relative data.

The maximum radar reflectivities correlated with optical observations of dark areas in 3 ways: (1) position of maximum radar reflectivity, (2) relative reflective rankings in the two frequencies, and (3) width of surface features. But despite these correlations, there were occasional discrepancies between reflectivity maxima and centers of dark areas (Sagan, Pollack, and Goldstein, 1966).

2.2.1.2 Polarization. — The thermal radio emission from a planet's disk has a net polarization governed by the surface dielectric constant, the roughness of the surface, the inclination of the line of sight, and the wavelength of the observation, according to Drake and Heiles (1964). They point out that if this polarization is measured in detail, it is possible to determine the following, including several surface parameters: (1) dielectric constant and hence some indication of surface constitution, (2) determination of whether limb darkening is due to atmospheric absorption or surface effects, (3) true surface temperature rather than equivalent blackbody disk temperature, and also diurnal, seasonal, and latitudinal temperature variations, (4) orientation of the rotation axis, (5) regions of increased roughness, especially those covered by extensive vegetation, and (6) the scale of roughness. Polarization measurements of Mars have an additional use, namely, to derive the atmosphere pressure.

To apply polarimetric methods, it is necessary to know the photometric properties of the surface (Hunten, 1966). A crude determination of atmospheric pressure can be made from the amount of molecular Rayleigh scattering as revealed by its characteristic polarization (Hunten, 1966), which is measured as a function of wavelength, position on the disk, and phase angle of Mars (Rea, 1965). Polarization measurements of the surface have shown that the properties of the top soil of the maria and continentes are similar (Öpik, 1966), and have helped to show that powdered limonite exists on the surface (Younkin, 1966; Loomis, 1965). During the last 9 apparitions of Mars, since 1948, Dollfus and Focas (1966b) made 5200 measurements of polarized light from Mars. They observed the bright and dark areas separately and studied the variation of polarization with wavelength from 0.45 to 1.05 microns. They also noted a seasonal modification of the microscopic texture of the soil in maria while it was darkening in the Martian spring (Dollfus and Focas, 1966b).

2.2.2 Visual Features

2.2.2.1 Polar Caps. — The polar caps are the most prominent visual features on the surface of Mars and they exhibit the most dramatic changes on a seasonal basis. Yet, they are computed to be only about 1 mm thick, consisting of finely divided crystals, resembling hoar frost (Rea, 1965). They are observed to migrate from one pole to the other (Goy, 1966), but they remain invisible between poles except for the darkening they produce in the maria. This suggests an evaporation of the material of the polar cap, which is observed as a shrinking of the cap in the local Martian spring (Miyamoto, 1966). The customary explanation of the darkening is that it is caused by the growth of plants, presumably made possible if the material from the polar cap is water vapor. Such a transport of water vapor has been confirmed spectroscopically (Spinrad, 1966a). One theory for the wave of darkening is that it is a nocturnal frost heave phenomenon, the moisture sublimating and progressing equatorward a little each day as water vapor (Otterman and Bronner, 1966). Another theory accounts for the wave of darkening, without invoking either plants or water vapor, as due to an increased wind in spring, which blows light colored dust off the darker looking surface of the maria (Sagan and Pollack, 1966a).

One would expect only a small amount of water substance in the form of water or ice on the surface of Mars in view of the small water vapor concentration in the atmosphere (Münch, 1966). If liquid water is present on the surface, it can only exist without evaporating where the total pressure is high enough to exceed the saturation vapor pressure of the water surface. The most likely water sites would be locations of low elevation in bright areas, where the surface pressure is high, perhaps 20 mb (Sagan and Pollack, 1966f). However, it is very improbable that liquid water could exist at all because the atmosphere is so dry that the ambient water vapor pressure is far less than the saturated value, as shown by the low H_2O abundances measured spectroscopically (Spinrad, 1966a; Öpik, 1966). Hence, there are no open bodies of liquid water and probably no moisture in the soil unless it is in the form of permafrost, which is reasonable if the temperature is below $0^{\circ}C$ to a depth of about 1 km (Öpik, 1966). Thin frost deposits could be expected to form over many areas in winter and then to sublimate to water vapor without going through the liquid phase (Öpik, 1966). However, the presence of water on Mars in the past is suggested by the observed erosion of craters (Sagan, 1966c).

The composition of the polar caps is not necessarily limited to H_2O ice, but is believed to consist of CO_2 ice or oxides of nitrogen (Leighton and Murray, 1966). The polymer, N_2O_4, of NO_2 has been proposed as the ingredient of the polar caps (Kiess, Karrer, and Kiess, 1960), but this suggestion was later dismissed (O'Leary, 1965). The case for dry ice in the polar caps (Leighton and Murray, 1966) has already been presented in the section on atmospheric pressure. Surprisingly enough, one model atmosphere gave results predicting an ice cap composed of solid CO_2 (Leovy, 1966b). It is possible that the total solid CO_2 on Mars may exceed the total CO_2 vapor in the atmosphere, the dry ice being located on the surface in the polar cap and under the surface as CO_2 permafrost (Leighton and Murray, 1966). In spite of the abundance of CO_2 on Mars, there can be no liquid CO_2 on or above the surface because the pressures are far below the triple-point pressure, as is even true in the Earth's atmosphere.

2.2.2.2 Craters. — The Mariner IV photographs showed that Mars has a heavily cratered surface. The statistics of Martian craters is similar to that of lunar craters, but their slopes seem to be greater than those of the Moon (Menzel, 1966). The frequency of craters on Mars was found to be greater than that of lunar maria, but less than that of lunar uplands (Leighton, 1966). The Mariner IV crater count extrapolated to 250,000 craters with diameters of at least 1.2 km, the diameter of the Arizona meteor crater (Öpik, 1966). The number of craters larger than a given diameter D is approximately proportional to D^{-2} (Öpik, 1966). The largest crater photographed was 120 km in diameter (Leighton, 1966). The surface appeared to have suffered little alteration during the history of Mars except for the impacts of objects, such as meteorites and asteroids, believed to have produced the craters (Sloan, 1966), of which craters with diameters larger than 20 km survived from the beginning (Öpik, 1966). The erosion rates for terrestrial deserts, the surface of Mars, and the surface of the Moon are in the proportion of 2000:70:1 (Öpik, 1966). Since the erosion rate is fastest for fresh craters, most of the Martian craters are strongly eroded with low walls (Öpik, 1966). One Mariner IV picture shows some bright spots and circular bright areas, believed to be frost on the surface, evidently associated with craters (Leighton, 1966).

Selected occurrences of white clouds may reveal the possible presence of volcanoes (Wells, 1966). Since the frequency distribution of Martian and lunar craters matches that of terrestrial volcanic craters, but not terrestrial meteoritic craters, this suggests a volcanic origin of most craters on Mars and the Moon (Simpson, 1966a; Simpson, 1966b). The alleged discrepancy with the frequency of meteor craters on the Earth may simply be due to the large mass of the terrestrial atmosphere, which shields the surface from most meteorites, thereby reducing the number of such craters even before erosion can begin. The volcanic theory of Martian craters is doubted because these craters do not resemble terrestrial volcanoes, and the thinness of the Martian atmosphere indicates a lack of past volcanic outgassing (Öpik, 1966).

<u>2.2.2.3 Canals</u>. — The canals of Mars are not as prominent as the polar caps, but they attracted more attention as possible artifacts of intelligent beings, an idea which is now obsolete. A comprehensive article on the Martian canals has been written by Sagan (1966b), which is summarized here. The present status of knowledge about canals is as follows: (1) they are features of bright areas, (2) the major canals are drawn in the same positions by independent observers, (3) they are hundreds to thousands of km long, (4) their widths are in dispute, (5) they are great circles, (6) they do not terminate in bright areas, (7) they terminate in either large dark areas or small circular dark areas known as "oases", (8) they undergo darkening in spring like the maria, and (9) they sometimes become double. Most of the suggestions for the interpretation of canals that have been offered by different investigators have run into difficulties. The hypotheses are as follows: (a) cracks in the crust of Mars formed as it cooled, (b) rays of lunar craters, (c) sequences of small craters accidently lined up or produced by the impact of secondary debris, (d) cracks from metor impacts that made dark oases, (e) rectilinear arrangement of sand dunes, (f) ridges lifted gravitationally by asteroids in near grazing collisions with Mars, (g) faults or escarpments, (h) ridges. Sagan (1966b) favors (h) because canals show up by radar Doppler spectroscopy as ridges with slopes exceeding 4° (Sagan, Pollack, and Goldstein, 1966). They are believed to be of tectonic origin resembling the Mid-Atlantic ridge with its seamounts and rift valley (Sagan, 1966b).

2.2.3 Topography

The Mariner IV occultation results showed height variations of the Martian surface of 3 miles (5 km) (Adelman, 1966). Yet, radar observations show that Mars is relatively flat (Goldstein and Gillmore, 1963). White clouds are indirect indications of areas of high surface elevation (Miyamoto, 1966; Wells, 1966). Dark maria (Wells, 1966) and canals (Sagan, 1966b) have been found by radar to be higher than the bright areas, based on the following principle. If the location from which specular backscattering comes is on the near side of a feature, it is a ridge; if on the far side, it is a depression (Sagan, Pollack, and Goldstein, 1966). The slopes of major dark areas generally exceed the slopes of areas undergoing secular changes (Pollack and Sagan, 1966). The former reach elevations of 10-20 km, while the latter have shallow slopes of 1-2^{o} and reach elevations of only about 6 km (Sagan, Pollack, and Goldstein, 1966). If all the dark areas are higher than the bright areas, the maria are the uplands and the continentes represent the lowlands. The S hemisphere includes most of the highlands, mainly between the equator and 30^{o}S latitude (bases on Loomis, 1965), and the N hemisphere has a large area of low flat plains (based on Rea, 1965). This arrangement is just the opposite of the Earth's topography, which has continental land masses primarily in the N hemisphere and ocean basins mostly in the S hemisphere.

2.2.4 Composition of Surface

To simulate the composition of the Martian surface in the laboratory, Draper and Adamcik (1966) prepared a mixture of minerals and studied its reflectivity. Iron oxide goethite and hematite in finely ground quartz and kaolin were too reflective to match Mars. Better results were obtained with two other models in which dark material magnetite blende was added (Draper and Adamcik, 1966). Very fine powders of diameters less than a few microns were hypothesized on Mars to explain the very low thermal conductivity of its surface (Leovy, 1966a), indicated by large diurnal temperature variations (Öpik, 1966). The conductivity of the surface was found to be comparable to that of air and less than that of terrestrial sand, which indicates a dusty unconsolidated top layer rather than a rocky surface (Öpik, 1966). Values of the dielectric constant on the Martian surface ranged from less than 1.8 to about 4 (Sagan and Pollack, 1966b). These values fit a wide range of inorganic materials provided they are pulverized or are in a highly porous form to a depth of at least several meters, porosities of 78 and 50% being required for the light and dark areas, respectively (Sagan and Pollack, 1966b). Limonite was the only mineral having the polarization, color, and albedo duplicating Mars (Younkin, 1966). Beneath the dust there is the possibility of a major H_2O ice layer, trapped as in permafrost (Gross, McGovern, and Rasool, 1966; Leighton and Murray, 1966; Strughold, 1966) by the prevailing low temperatures.

3. VENUS

3.1 DIURNAL ROTATION

The meteorology of Venus depends upon its axial rotation rate and upon the orientation of its axis of rotation. Consequently, these two topics will be discussed before the Venusian atmosphere is considered.

3.1.1 Period and Sense of Rotation

The period of axial rotation is important in that it determines the nature and diurnal variations of the circulation pattern and the temperature distribution of the atmosphere of Venus.

Moore (1966) made studies of vague visible shadings on Venus for 30 years, but found that they contributed no information about the rotation period. Radar observations show that the period of Venus is remarkably slow. It is of the same order of magnitude as the period of orbital revolution (Thomson, 1966), suggesting that Venus is in synchronous rotation, keeping the same face pointing toward the sun. But sun-synchronous rotation is ruled out by microwave observations, which showed a phase effect on brightness temperature indicating a retrograde rotation (Drake, 1966), based on the choice of the terminator with the higher temperature as being the line marking sunset. The retrograde rotation was also indicated at inferior conjunction by a minimum in the limb-to-limb Doppler spread in radar frequency (Dyce, 1966), synonymous with a minimum apparent rate of rotation, relative to an observer on the Earth, at that time.

By careful radar studies, Goldstein (1966b) deduced from depolarized spectrograms a retrograde rotation period of 242.6 ± 0.6 days. Within the limit of probable error, this includes an Earth-synchronous period of 243.16 days retrograde, defined such that there are just 5 days and nights on Venus between each inferior conjunction (Goldstein, 1966b), 584 days apart, giving a complete day on Venus equal to 117 terrestrial days. Whatever axis, or face, of Venus is oriented toward the Earth at one inferior conjunction is oriented that way at all inferior conjunctions if the rotation period is exactly Earth-synchronous, according to Goldreich and Peale (1966). They found that the gravitational effect of the Earth on Venus was too small to make the rotation of Venus synchronous with the Earth unless Venus has a very much larger axial asymmetry than the Earth. Radar determinations of the locations of previously identified topographical features near the time of the next inferior conjunction will provide the most accurate measure of the period of Venusian axial rotation yet possible (Goldstein, 1966b). Since Venus has a much slower rotation than the Earth, it may have a much smaller magnetic moment than that of the Earth (Dickerman, 1966), which is also indicated by the Mariner II observations.

3.1.2 Axis of Rotation

The orientation of the rotational axis of Venus relative to the plane of its orbit is a significant parameter in seasonal meteorological variations on Venus. Since a season lasts 56 terrestrial days (1/4 of 225-day Venus year), it is almost of the same duration as daylight, or a night, of $58\frac{1}{2}$ terrestrial days (1/2 of 117-day Venus day). Since the seasonal and diurnal effects are on the same time scale, they should be considered together. At one point on Venus, an annual sequence might run like this: spring night, summer day, autumn night, then winter day; or the sun reaches the solstices during the daytime and the equinoxes at night. The seasonal effect is produced by the difference in the elevation angles of the sun on the two days. At a point on the opposite side of Venus, the solstices would be at night and the equinoxes would occur during the day. Since the sun would be at the same elevation angles on the two days, there would be no seasonal effect there. Consequently, seasonal effects on Venus should occur predominantly on one side of Venus. Since the length of a season is about 4 percent shorter than a half day, the region of maximum seasonal effect should not remain fixed, but should drift slowly westward in longitude on Venus, and return to its original position after about 8 terrestrial years.

The spin axis of Venus is nearly perpendicular to the orbital plane, being within 10° of the pole of the orbit according to Carpenter (1966), who gives the N pole of the rotation axis at a right ascension of $255^{\circ} + 10^{\circ} - 4^{\circ}$ and declination of $68^{\circ} \pm 4^{\circ}$. Goldstein (1966b) gives a different spin axis orientation placing the S pole at right ascension $98^{\circ} \pm 5^{\circ}$ and declination $-69^{\circ} \pm 2^{\circ}$ (equivalent to N pole at $278^{\circ} \pm 5^{\circ}$ right ascension), which is even closer to the orbital pole. The tilt of the rotation axis may be only 6° to the orbital axis (Dyce, 1966). Consequently the seasonal variations on Venus between winter and summer are of small magnitude, certainly less than the changes between day and night (except at the poles of Venus, where spring and summer correspond to daylight; and fall and winter, night; the same as at the terrestrial poles).

3.2 ATMOSPHERE

Important observations of the atmosphere of Venus were made in July, 1959, when the planet occulted the first magnitude star, Regulus, and again in December, 1962 by Mariner II in its fly-by. The resulting gain of knowledge pertained to the perpetual clouds and the atmosphere above the clouds. The mystery of the atmosphere below the clouds and of the surface still remains, although significant inroads toward dispelling this mystery have been made by microwave and radar penetrations of the shroud. Proper interpretations of the data await the developments of more comprehensive theories than are now available.

3.2.1 Appearance of Venus

3.2.1.1 Albedo. —

3.2.1.1.1 Variation with Wavelength: One of the outstanding characteristics of Venus, in contrast to the other planets, is its high albedo in the visible spectrum. Its brightness is caused by the strong reflection of its perpetual cloud layer. Sinton (1963) presented a table (see below) of average albedos for wavelength intervals of the solar spectrum, from which he derived an overall weighted average visual albedo of 0.734 for Venus. There is a decrease of albedo toward shorter wavelengths of the ultraviolet and also toward longer infrared wavelengths. Between 0.4 and 1.3 microns the albedo has a high and nearly constant value.

TABLE 3

VENUS ALBEDO VERSUS WAVELENGTH

Wavelength λ (microns)	Fraction of Solar Energy f	Venus Albedo A	Product fA
0 - .35	.070	(.40)	.028
.35 - .40	.053	.53	.028
.40 - .50	.126	.78	.098
.50 - .60	.129	.87	.112
.60 - 1.2	.428	(.87)	.372
1.2 - 1.5	.074	(.75)	.056
1.5 - 1.9	.053	(.52)	.028
1.9 - 2.7	.040	(.40)	.016
2.7 and more	.027	(.15)	.004
TOTAL	1.000		$\bar{A}$ = .734

The relatively low albedo in the blue spectral region may result from cloud absorption according to Öpik (Dickerman, 1966). The ultraviolet is characterized by a broad absorption-like feature beginning at 3300A and reaching a reflectivity minimum at 2500A, which can be interpreted in one of two ways (Evans, 1966): (1) the combined effects of Rayleigh scattering from a 60-mb atmosphere above a cloud surface which has a reflectivity decreasing toward the ultraviolet, or (2) Rayleigh scattering in an atmosphere with ozone, combined with a wavelength independent reflectivity from a cloud surface.

The decreased albedo in the infrared may represent a lower intrinsic reflectivity of the clouds (Sagan and Pollack, 1966c). At the longer wavelengths the single scattering albedo decreases and it becomes the dominant factor in the albedo (Sagan and Pollack, 1966c) because the multiple scattering decreases faster and becomes negligible. The albedo variation in the visible and near infrared spectral regions can be reproduced theoretically if allowance is made for gaseous absorption (Sagan and Pollack, 1966d) and if the Mie scattering theory given by a computer solution is used (Sagan and Pollack, 1965b). When the ultraviolet and visible albedo values are used along with infrared reflectivity measurements, an effective temperature at the cloudtop level can be derived (Sinton, 1964).

3.2.1.1.2 Markings: In a telescope, Venus generally appears as a featureless disk, but indistinct markings have sometimes been detected in the ultraviolet and infrared spectral regions. Smith (1963) looked for ground observations of any Venus markings which might have been noted in December, 1962, for comparison with observations made by Mariner II while it was in its close approach to the planet. A summary of the report by Smith (1963) is presented below.

Under average conditions, ultraviolet photographs usually reveal a series of light and dark bands, the "Ross bands", which are somewhat parallel to each other, but together are normal to the line joining the cusps. The intensity, number, and inclination of the bands exhibit noticeable changes from day to day. The Ross bands are not visible in the ultraviolet at large phase angles, when Venus is near inferior conjunction, nor can they be seen in the visible spectrum at any phase angle of Venus. In the near infrared, dusky markings of extremely low contrast are infrequently noted, the typical near infrared image being featureless. During the Mariner II encounter, Venus was essentially featureless, which is normal for its fairly large phase angle of that time. Markings were detected on only 2 days, in both cases only in the ultraviolet, where weak Ross bands were detected.

3.2.1.1.3 Limb Darkening: The meaning of limb darkening is that the brightness decreases with an increase in the angular distance of the direction of viewing from the center of a planet's disk. (For a more detailed statement, see the section on limb darkening and brightening for Mars.) Pollack and Sagan (1965b) prepared theoretical models of the Venusian atmosphere and derived the corresponding profiles of infrared limb darkening. Then they compared their theoretical profiles with observed profiles (Sagan and Pollack, 1965a) to determine the relative validities of their models. These two papers are sketched below.

Three general categories of models of the atmosphere and clouds of Venus were invented as follows (Pollack and Sagan, 1965b). The Models A attributed the limb darkening to the temperature lapse rate in an unknown approximately grey absorber above the cloud layer. Other properties required of the absorber made these models rather unrealistic. In Models B, the limb darkening was attributed partly to the lapse rate and partly to the angular dependence of emissivity of approximately grey multiply scattering clouds. Model B1, based on convective equilibrium, gave a theoretical limb darkening which would fit the observations for a moderate value of the albedo for infrared single scattering by diffuse clouds. Model B2, based on radiative equilibrium, would account for the limb darkening if there was a moderate albedo and if the 8 to 12 micron opacity was about 2½ times the mean opacity, a property of silicate clouds. The Models C hypothesized clouds which coherently redistribute radiation emitted by the underlying atmosphere. The absence of any marked spectral features in the 8 to 13 micron Venus spectrum lead to the rejection of these Models C. The observed limb darkening (Pollack and Sagan, 1965b) was such that the 8 to 13 micron brightness was directly proportional to the square root of the cosine of the zenith angle of the spectrophotometer as it would be viewed from Venus. Each of Models A and B could be made to fit the observations if appropriate atmospheric and cloud parameters were specified (Sagan and Pollack, 1965a).

In the microwave region of the spectrum, limb darkening was likewise observed and attributed to a variation of surface emissivity with angle and to slant path attenuation by microwave absorbers in the atmosphere and clouds(Sagan and Pollack, 1965b). The observations can be best explained by an absorbing cloud or scatterers which are arbitrarily rather than continuously distributed. Microwave limb darkening indicates that the source of the radiative emission is at or near the surface of Venus rather than at the ionosphere, which would produce limb brightening (Dickerman, 1966). Mariner II microwave observations, at centimeter wavelengths, of limb darkening indicated high thermal emission from the surface (Kaplan, 1964; Barath, Barrett, Jones, and Lilley, 1964). Along a diameter of the disk of Venus which is perpendicular to the polarization of a radio telescope there is limb darkening, but in the orthogonal direction, there is enhanced emission (as the limb is approached), a maximum emission at the Brewster angle, and then limb darkening (Drake and Heiles, 1964).

3.2.1.2 Cusp Caps. — Cusp caps are optical enlargements or extensions of the horns of the Venus crescent seen near inferior conjunction. They are a manifestation of the phase anomaly, which refers to the excess of the difference in Venus longitude between its visible bright limb and terminator as compared to the geometrically predicted longitude difference.

An observed asymmetry in the angular lengths of the 2 cusp caps is likewise an asymmetry about the orbital radius vector of Venus, which indicates that Venus has a non-synchronous rotation (Smith and Baines, 1964). Hypotheses suggested for the cusp caps include ionized clouds attracted to the magnetic poles of Venus (Pither, 1962) and auroras indicated by an alleged correlation with solar flares (Pither, 1963). The fact that cusp caps do not cover a portion of the dark face of Venus away from the terminator is observational evidence that they are not auroras, but instead could be attributed to clouds at high altitudes (Pearce, 1964). The size of the cusp caps increases with decreasing wavelength from red to blue, as viewed through colored filters (Pither, 1962). This is in accord with observations which showed that Venus is brighter and bluer than predicted when its phase angle is greater than 130^{o} (de Vaucouleurs, 1961). The brightness of the planet is too great to be explained by scattering by gases, but could be produced by scattering from spherical particles with radii 0.3 to 0.5 microns and with a refractive index of 1.33 in a layer not more than a few kilometers thick (de Vaucouleurs, 1961).

At inferior conjunction, when Venus appears closest to the sun, the cusps have been observed to join, forming a complete ring around the dark face of the planet (de Vaucouleurs, 1961). This phenomenon has been explained by the refraction and multiple scattering of solar radiation into the geometric shadow cone of Venus (Moore and Schilling, 1965). Measurements of phase anomaly have been made (Nicolson and Moore, 1964), particularly near the time of dichotomy, when half the disk of Venus is illuminated, which is very close to the time of maximum elongation of the planet from the sun. A discrepancy between the times of observed and theoretical dichotomy amounting to 4 to 6 days has been explained mathematically (Warner, 1963). The observations show that the observed phase angle of Venus is less, or the fraction of the disk illuminated is always larger, than required by geometry. A model to account for this is an atmosphere with 3 cloud layers, of which the highest, several tens of km above the middle layer, is responsible for the extension of the horns of Venus (Deirmendjian, 1966).

3.2.1.3 Ashen Light. — When Venus is a crescent, its dark face can sometimes be seen shining faintly, a phenomenon known as "ashen light". It constitutes an emission of light rather than a reflection of sunlight, which accounts for the rest of the light of Venus. Electrical hypotheses have been offered to explain the emission. One possibility which is supported spectroscopically, attributes ashen light to atmospheric electrical discharges analogous to terrestrial aurorae (Moore, 1966). In the absence of a magnetic field around Venus, the auroras would not be restricted to high latitudes, but also they would not be concentrated in location and therefore might be too weak to be detected visually. Another possibility is that there are glow discharges in the atmosphere of Venus similar to those produced by CO_2 and SO_2 in the laboratory, which yield X-band microwave signals, a possible contribution to the microwave brightness of Venus (Applebaum, Harteck, Reeves, and Thompson, 1966).

3.2.2 Composition of Atmosphere

3.2.2.1 Gases and Vapors. —

3.2.2.1.1 Retention of Atmosphere: The similarity of Venus to the Earth with regard to its mass and size gives it only slightly lower values of the acceleration of gravity and escape velocity for atmospheric gases. Consequently, Venus is more able than Mars to retain an atmosphere with gases as light as those of the Earth's atmosphere.

Because of the relative absence of oxygen on Venus, photodissociation of water may be very efficient and could have conceivably led to the escape of amounts of water comparable to the discrepancy between the large amounts of water on Earth and its scarcity on Venus (Sagan, 1965).

3.2.2.1.2 Hydrogen: One of the products of the photodissociation of water is hydrogen. This lightest of all gases is able to escape from the atmosphere of Venus as it does from the Earth's atmosphere (Dickerman, 1966). Due to diffusion, its concentration relative to that of other gases will be greatest at the top of the atmosphere, but even there its abundance is considered to be negligible.

3.2.2.1.3 Hydrocarbon: The possible existence of traces of CH_2O (formaldehyde) has been indicated (Prokof'yev, 1965) by a broad emission feature in the visible spectrum, but it has not been confirmed (Dickerman, 1966). Negative results have been obtained in the search for CH_4(methane), C_2H_4, and C_2H_6 (ethane) (Dickerman, 1966). Likewise no other hydrocarbons have been detected in the atmosphere of Venus.

3.2.2.1.4 Carbon Oxides: The first constituent of the Venusian atmosphere detected was CO_2, observed by Adams and Dunham in 1932 (Dickerman, 1966). There has been considerable uncertainty about the relative atmospheric concentration of CO_2, estimates varying from as high as 90 percent (Van Tassel and Salisbury, 1965) to as low as 4 percent (Spinrad, 1963). The latter is based on measurements of the absorption in the 7820A spectral band of CO_2. At present, the estimated relative concentration is 2 to 10 percent (Dickerman, 1966). If the absorption bands of CO_2 are produced in the optically thick atmosphere of Venus through multiple scatterings, knowledge of the atmosphere is much less precise than commonly believed (Chamberlain, 1965). The upper limit of CO_2 abundance is of the order of magnitude of 10^4 m atm (Sagan and Pollack, 1966d), but the abundance in the visible cloud layer may be only 10^2 m atm (Prokof'yev, 1965). Large amounts of carbon on Venus would appear in atmospheric CO_2 instead of carbonaceous rock (Dickerman, 1966). Consequently, if Venus and the Earth originally had about the same amount of CO_2, it is reasonable to find that the CO_2 content (perhaps 5 percent by volume) of the Venusian atmosphere is approximately equal to the CO_2 in carbonates in the crust of the Earth (Sagan, 1965). The observed CO_2 abundance in the atmosphere of Venus is of the order of magnitude corresponding to the equilibrium vapor pressure of CO_2 expected at the temperatures of Venus for reactions among carbonates, carbon dioxide, and silicates (Mueller,1964).

Several weak absorption bands of CO_2 have been detected, including the "hot band" at 8736 to 8768A, which is mainly the rotational fine structure of an excited vibrational transition of CO_2 (Spinrad, 1966b). Paradoxically, the weak bands have indicated different CO_2 abundances (Chamberlain, 1965). CO_2 has numerous spectral lines, such as at 7158, 7828, 7891, 8698A,and between 1 and 10 microns, indicating abundances of 10^2 to 10^3 m-atm (Dickerman, 1966). Some carbon dioxide has been found consisting of carbon or oxygen isotopes, with abundances in about the same ratios as on the Earth (Chamberlain, 1965). The spectral lines of $C^{13}O_2$ at 1.475 microns and $C^{12}O^{16}O^{18}$ at 2.15 microns have been identified by several observers.

The other important oxide of carbon, CO, is expected to be present in the CO_2-rich Venus atmosphere (Sagan, 1965), since it is produced photochemically from CO_2 in the upper atmosphere, but it may combine with atomic oxygen to form CO_2 again (Shimizu, 1966). Carbon monoxide has been tentatively identified from its spectral band at 2.345 microns (Dickerman, 1966). The CO abundance is uncertain, but it may be about 2 cm-atm (Shimizu, 1966) or as high as about 23 cm-atm (Sinton, 1963).

3.2.2.1.5 Nitrogen: Nitrogen is presumed to be the principal constituent of the atmosphere of Venus, although its identification is by default, since the nitrogen absorption lines are located in presently inaccessible parts of the electromagnetic spectrum (Sagan, 1965). The abundance of N_2 has been assumed to be as low as 9 percent, but this was in an atmosphere which was presumed to be mostly of CO_2 (Van Tassel and Salisbury, 1965). An unconfirmed identification of N_2^+ in the 4300A region has been made in connection with emission from the night sky (Dickerman, 1966). In the high atmosphere of Venus, collisions between N_2 and CO_2 are assumed to be spectroscopically important (Dickerman, 1966). While atomic nitrogen has been detected, there are no quantitative data on its amount (Prokof'yev, 1965).

3.2.2.1.6 Nitrogen Oxides: The predicted equilibrium abundances of nitrogen oxides are extremely low on all the planets (Lippincott, Eck, Dayhoff, and Sagan, 1966). Negative results were obtained from searches for N_2O in the atmosphere of Venus (Dickerman, 1966). The nitrogen compound, NH_2 (ammonia),has not been detected either.

3.2.2.1.7 Oxygen and Ozone: Venus has been characterized as having an oxygen-free atmosphere, there being no reliable identification of O_2 there (Sagan, 1965). However, oxygen has been tentatively identified on Venus (Plummer and Strong, 1965; Prokof'yev and Petrova, 1964). An estimated upper limit of abundance has been set at 80 m-atm (Dickerman, 1966). This upper limit has been refined to 57cm-atm, or smaller than terrestrial abundance by a factor of 2800, on the basis of a lack of Doppler-displaced Venus O_2 lines in the telluric A-band (Spinrad and Richarson, 1965). The unexpectedly high absorption of microwaves at 8.6 mm may be due to oxygen rather than water vapor (Gibson and Corbett, 1965). The effect of diffusion should be taken into account in the determination of the vertical distribution of O_2 molecules (Shimizu, 1966). Oxygen may be produced by the photodissociation of water, but not by photosynthesis of plants because surface temperatures are probably too high to permit life (Sagan, 1965).

Atomic oxygen, presumably present in the upper atmosphere of Venus, is produced by the photodissociation of CO_2, but is lost if it combines with CO to form CO_2 again (Shimizu, 1966). There is no reliable information about the abundance of atomic oxygen.

Ozone may be responsible for the slight yellow tint of Venus because of its absorption of ultraviolet radiation (Plummer and Strong, 1965), which slightly reduces the intensity of blue, the complimentary color of yellow. Evans (1966) prepared a model of the atmosphere of Venus containing 1/3 to 1/10 the amount of terrestrial O_3. In his model, the level of maximum O_3 would be very close to the height of the upper surface of the reflecting cloud layer, which would be near the 230-mb pressure altitude (Evans, 1966). The abundance of O_3 suggests a source other than CO_2:N_2 photochemistry, possibly C_2:N_2:H_2O (Evans, 1966). Another O_3 source is the combination of molecular and atomic oxygen in the high atmosphere, but abundances may be quite low.

3.2.2.1.8 Water Vapor: Water vapor has attracted more attention than any other constituent of the atmosphere of Venus, yet the infrared spectrum indicates only small amounts present there (Menzel, 1966). The 1.13 micron band is useful for the identification of water vapor because liquid water and ice have no strong absorption at that wavelength (Deirmendjian, 1966). From absorption in this band, Bottema, Plummer, and Strong (1964) derived amounts of water vapor equal to 222 microns and 52 microns of precipitable water for pressures of 90 mb and 600 mb, respectively. Deduced Venusian water vapor mixing ratios are actually greater than those observed in the terrestrial stratosphere above 12 km, but there is some uncertainty about the level in the Venus atmosphere at which the 1.13 micron bands form (Deirmendjian, 1966). Menzel and Whipple calculated a precipitable water content of 130 microns for the vapor above the clouds. Dollfus (1966c) found from measurements at a wavelength of 1.40 microns about 100 microns of precipitable water. A vertical distribution of humidity prepared by Dollfus (1966c) is presented in the following table for Venus.

TABLE 4

DISTRIBUTION OF WATER VAPOR ABOVE MAIN CLOUD LAYER OF VENUS

Layer of Air Above Clouds (km)	Quantity of Water Vapor for Saturation in the Layer (gm/cm^2)	Humidity Estimated (%)	Quantity of Water Vapor in the Layer (gm/cm^2)
0.0-0.5	0.0060	100	0.0060
0.5-1.0	0.0040	60	0.0024
1.0-2.0	0.0040	30	0.0012
2.0-3.0	0.0030	20	0.0006
3.0-7.0	(0.0040)	20	0.0008
Above 7.0	------	---	(0.0005)
Total Quantity of Water Vapor Above the Clouds: (or 115 microns of precipitable water)			0.0115 gm/cm^2

Another spectroscopic determination gave an extreme upper limit of 125 microns of precipitable water based on a preliminary estimate of absorption in the red wing of one of the faint water vapor spectral lines near 8200A (Belton and Hunten, 1966b). Another measurement of the same lines yielded a precipitable water content of 60 microns (Spinrad and Shawl, 1966). These faint lines, long sought through photographic methods, are of crucial importance to the interpretation of the stronger and perhaps saturated water bands at 1.13 and 1.4 microns (Belton and Hunten, 1966b).

The question remains as to how much water vapor there is below the upper surface of the clouds. The total precipitable water content of the atmosphere is less than 1 meter (Sagan and Pollack, 1966d) and probably no more than a few tens of centimeters (Drake, 1966). Öpik assumes that there is an insignficant amount of water vapor in the lower atmosphere (Dickerman, 1966) in spite of the high potential amounts at the high Venus temperatures. Moreover, Spinrad found that water vapor can account for the microwave spectrum of Venus only if its abundance is several orders of magnitude higher than indicated by infrared studies (Ho, Kaufman, and Thaddeus, 1966).

3.2.2.1.9 Neon and Argon: It is very difficult to determine the presence and abundance of any inert gas because of its lack of a detectable spectrum. Sometimes neon and argon are assumed to be present in a model atmosphere of Venus, such as by Ho, Kaufman, and Thaddeus (1966). A quantitative model by Van Tassel and Salisbury (1965) assigned a relative abundance of 1 percent to argon and gases other than nitrogen and carbon dioxide. The abundance of neon would be a lot higher than assumed in this model if the gases of the Venusian atmosphere had the same relative abundances as the sun corrected for the escape of hydrogen and helium for Venus. The two chief constituents of the latter are neon and carbon dioxide, the number of neon atoms being between 5 and 25 for every 10 molecules of CO_2 (Suess, 1964). On this same scale, there would be only 1.5 nitrogen molecules and 0.15 argon atoms. Consequently, neon could be the most abundant gas of the Venusian atmosphere instead of nitrogen.

3.2.2.1.10 Sulfur Compounds: The abundances of the vapors of major sulfur compounds in the lower atmosphere are assumed to be determined by an equilibrium between their vapor pressures and those of the minerals, $CaSO_4$, $CaSiO_3$, SiO_2, Fe_3O_4, FeS, and FeS_2 (Mueller, 1965). The amounts of sulfur compounds in the atmosphere are assumed to be negligible since their spectral bands have not been detected.

3.2.2.1.11 Summary: Only 4 gases are expected to constitute 99.99 percent of the Venusian atmosphere: nitrogen, neon, carbon dioxide, and water vapor. Together, nitrogen and neon may constitute about 94½ percent by volume, of which perhaps 63 percent is N_2 and 31½ percent is Ne (based on the choice of the lower limit of abundance). CO_2 and H_2O constitute about 5 percent and ½ percent, respectively. The residual .01 percent includes all other gases and vapors, such as hydrogen, hydrocarbons, carbon monoxide, ammonia, nitrogen oxides, oxygen, ozone, and sulfur compounds. (Note: the relative abundances are not known nearly as precisely as implied by the percentage values, which were computed to the nearest ½ percent to fit 5% CO_2, $\%N_2/\%Ne = 2$, $\%CO_2/\%H_2O = 10$, and sum = 100%.)

3.2.2.2 Suspensions. —

3.2.2.2.1 Clouds: The generally high albedo of Venus implies it has more compact clouds than those on the Earth (Pollack and Sagan, 1965b). The Venusian atmosphere appears to contain large amounts of particulate matter which seems to be permanently in suspension and may be arranged in 3 distinct layers (Deirmendjian, 1966). Clouds are composed partly of particles with a micron size range (Dickerman, 1966). Lyot reported in 1929 that he found particles 2 to 3 microns in diameter from a polarimetric study of clouds (Dollfus, 1966c). The top of the principal cloud surface would be at a level where the pressure is less than 1000 mb, which is based on an observed constant reflectivity near 3500A (Evans, 1966). This cloud-top pressure was estimated by Sagan and Kellogg to be between 600 mb on the illuminated side of Venus and 90 mb on its dark side, and the cloud-top altitudes were located between 60 and 100 km (Dickerman, 1966). The observed small decrease of infrared temperature from the light to the dark side of Venus may be due to the release of latent heat, an indication of cloud formation by the condensation or sublimation of water vapor (Strong, 1966). This leads to clouds of water drops or ice crystals, respectively, which will be discussed in the next two sections.

3.2.2.2.2 Water Drops: If the water vapor mixing ratio is assumed to increase with decreased altitude at a rate comparable to that in the Earth's upper troposphere, for a realistic water vapor content above the cloud layer there will be enough water vapor at the cloud top to approximate saturation, which lends credulity to the hypothesis that the clouds are aqueous (Ohring, 1966). Such clouds form in a saturated atmosphere in the presence of condensation nuclei (Sagan and Pollack, 1966c). Observations of microwave radiation from Venus place some requirements on cloud composition. The marked difference between the brightness temperatures at wavelengths of 13.5 and 19 mm cannot be due to uncondensed water vapor, but requires molecules that condense or polymerize into a liquid at a temperature of at least $350^{o}K$ and also produce spectral lines at wavelengths near and less than 13.5 mm (Jones, 1966). Liquid droplets of water or of other polar molecules with an abundance of at least a fraction of a gm/cm^2 column are found to have sufficient attenuation to account for the observed opacity of the clouds (Ho, Kaufman, and Thaddeus, 1966). However, infrared spectroscopic observations require very low upper limits for the abundance of polar molecules, such as water vapor, suggesting that, without water saturation, an unknown constituent of the atmosphere may be responsible for the microwave opacity (Ho, Kaufman, and Thaddeus, 1966). On the other hand, radar astronomical observations call for 0.1 to 0.2 g/cm^2 of aerosols of a polar liquid, which would correspond to water cloud heights of only 1 to 2 kilometers (Kuzmin, 1966), insufficiently high to have the low temperatures measured in the infrared. The relationship between temperature and wavelength may be satisfied by supercooled water (Basharinov and Kutuza, 1965) or some hydrocarbons such as CH_3OH, C_6H_5Cl, etc. (Kuzmin, 1966).

3.2.2.2.3 Ice Crystals: Ice crystals were discovered in the atmosphere of Venus by its infrared spectrum between 1.7 and 3.4 microns, which matched that of ice (Bottema, Plummer, Strong, and Zander, 1964), but not that of terrestrial water clouds (Deirmendjian, 1966). Another indication of ice crystals was the tentative detection of a 22° halo effect when Venus was near a phase angle of 158° (O'Leary, 1966). Deirmendjian's 3-layer model consists of a lower cloud layer of optical thickness of the order of 10^2 consisting of water drops, some being as big as raindrops; an intermediate ice cloud layer with a thickness of 10^{-1} to 1 located a few km higher; and a superior layer of an unknown constituent with a thickness of 10^{-3}, several tens of km higher. This model does not require hailstone-sized particles to explain the microwave temperature differences on Venus (Deirmendjian, 1966). However, the water and ice do not have to be confined to separate layers. The principal reflecting cloud sheet is believed by Sagan and Pollack (1966d) to be composed primarily of ice crystals and secondarily of some water drops near its base. To account for the attenuation in the microwave spectrum at mm wavelengths, they proposed absorption by 0.1 gm/cm^2 of supercooled water just above the cloud base (Sagan and Pollack, 1966c). This attenuation can also be explained by absorption due to carbon dioxide, dust, or ice crystals (Drake, 1966). The near infrared spectrum showed a very impressive fit to ice except at a wavelength of 2 microns, where there was some abosrption, mostly by CO_2, according to Sagan and Pollack(1966c). From several infrared determinations they derived the mean particle radius to be between 7.5 and 10 microns. The mutual agreement of the results along with thermometric determinations of temperature well below 0°C support the identification of ice as the principal constituent of the Venusian clouds, but not water, which differs significantly from ice in its values of the extinction coefficient as a function of wavelength (Sagan and Pollack, 1966c). The maxima and minima of scattering occur at slightly longer wavelengths for ice than for water (Sagan and Pollack, 1966c). Besides liquid water, hydrocarbons and dust were found to be unlikely as a principal constituent of the top of the main reflecting cloud layer (Sagan and Pollack, 1966e).

3.2.2.2.4 Haze Layer: The only haze that can be optically detected in the atmosphere of Venus must lie above the visible cloud layer. Dollfus(1966c) proposed 4 hypotheses for the vertical distribution of aerosol particles of about 1.5 microns in diameter: the number density (1) decreasing with increasing height in proportion to the pressure, (2) decreasing twice as rapidly as pressure, (3) remaining constant up to a height of 10 km and being zero above 10 km, or (4) being constant up to 7.5 km and zero above that. Hypothesis (2) gave the best agreement with photometric observations (Dollfus, 1966c).

3.2.2.2.5 Dust Layer: Since the composition of clouds shrouding Venus is controversial, it may include water vapor and dust as possible constituents (VanTassel and Salisbury, 1965). Dust, if present, must be tenuously distributed through a massive atmosphere, as suggested by the origin of near infrared CO_2 bands at relatively high pressures, or low altitudes (Ho, Kaufman, and Thaddeus, 1966). The opacity of dust theoretically accounts for the amplitude of 8 mm microwave brightness temperature variations with the phase of Venus, whereas the theoretical opacities of water vapor and CO_2 conflict with the observations (Sagan and Pollack, 1965b). To provide simultaneously for high microwave opacity and low abundance of polar molecules for the infrared, one could choose volcanic ash, an aerosol which is often observed to be hygroscopic (Ho, Kaufman, and Thaddeus, 1966). While it supplies the high dielectric loss characteristic of a liquid droplet, it removes water vapor from the atmosphere (Ho, Kaufman, and Thaddeus, 1966), but its hygroscopic action is probably too weak to reduce the water vapor content of the Venusian atmosphere enough to satisfy infrared requirement. Opik(1961) introduced dust as the essential ingredient of his aeolosphere theory, which will be discussed in the section on the heat budget.

3.2.3 General Circulation

As there are no satisfactory observations of cloud motions in the Venusian atmosphere, the general circulation must be reconstructed from models, which are necessarily simpler than the real circulation. In some models winds are hypothesized to provide the force necessary to raise solid aerosols, such as dust, off the surface. In a model with a surface pressure in tens or hundreds of atmospheres, the theoretical windspeeds are very low (Ho, Kaufman, and Thaddeus, 1966), but the atmospheric density would be so high that even these light winds could lift dust into the atmosphere. In 4 models of the circulation of Venus, the mean windspeeds were found to be 2 to 8 m/sec (Ohring, Tang, and Mariano, 1965).

The atmospheric circulation derives its energy from solar radiation. Venus rotates so slowly about its axis that its general circulation is not organized like the terrestrial circulation in latitude zones parallel to the equator. Instead, the circulation of the Venusian atmosphere is bipolar with respect to the subsolar and antisolar points, where the atmosphere is convectively rising and sinking, respectively (Plummer and Strong, 1965). If the atmosphere contains a condensable cloud generating substance, this simple large scale circulation pattern requires the whole sunlit hemisphere to be cloud covered, as observed, and the dark hemisphere to be clear, contrary to observation (Opik, 1963). The dark-side discrepancy can be eliminated by the assumption of a large descending dust layer extending to the surface (Opik, 1963). Another model calls for a cloud cover over most of the dark side as well as over all of the sunlit side of Venus on the basis of widespread small updrafts (Goody and Robinson, 1966) and a limited region of downdraft around the antisolar point to satisfy continuity requirements. The Coriolis forces are strong enough to have a secondary influence on the flow in spite of the slow rotation rate of Venus (Goody and Robinson, 1966). Ultraviolet cloud patterns suggested to Mintz (1961b) that the general circulation consisted of large low-level horizontal cells transporting heat from the day side to the night side of Venus, surmounted by a high circumpolar vortex. In view of more recent knowledge about the slowness of the planet's rotation, a circumpolar vortex appears to be asymmetrical. It should extend to a lower latitude on the side of the antisolar point since it should encircle the coldest region of the lower atmosphere.

3.2.4 Temperature Distribution

3.2.4.1 Temperature. —

3.2.4.1.1 Surface Temperature: Surface temperatures on Venus have stirred up a great deal of interest because of their surprisingly high values, as revealed by measurements of microwave emission. The effective brightness temperatures of 600 to 700°K at radio wavelengths of 3 to 10 cm (Barrett, 1965) are believed to approximate the thermometric surface temperatures (Drake, 1966). Microwave radiation is interpreted as originating at the surface since it shows no rapid fluctuations, no appreciable polarization, and is nearly independent of frequency over more than one order of magnitude: 0.4 to 10 Gc/S (75 to 3 cm) (Ho, Kaufman, and Thaddeus, 1966). Menzel(1966) commented that one always hopes that some other interpretation can be found for the observed high brightness temperatures. The lack of realistic models to account for high surface temperatures has led to a search for alternative theories, such as the possibility of glow discharge (Applebaum, Harteck, Reeves, and Thompson, 1966). Plummer and Strong(1966) hypothesized that some of the microwave radiation is of nonthermal origin. At the microwave wavelengths, which is on the long-wavelength tail of the blackbody curve, it is permissible to make the approximation given by the Rayleigh-Jeans law, namely, that the emission in a small spectral interval is proportional to the absolute temperature. Plummer and Strong(1966) found that nonthermal emission may account for 3/10 of the total microwave emission, leaving 7/10 of the emission as thermal emission from the surface. The average absolute brightness temperature of 560°K observed for the dark hemisphere was attributed to a surface temperature of 392°K and to nonthermal radiation (equivalent to 168°K) from the clouds. Similarly, an average surface temperature of 238°C(511°K) was found for the bright hemisphere (Plummer and Strong, 1966). But these temperature values are not generally accepted because of uncertainty about the amounts of nonthermal emission.

The microwave brightness temperatures vary with the wavelength (Figure 7). They decrease toward longer wavelengths from about 600°K at 20 cm to about 500°K at 68 cm for an unknown reason (Drake, 1966). The brightness temperatures decrease even more toward shorter wavelengths from about 600°K at 2 cm to only 350°K at 8 mm and 300°K at 1.3 mm due to attenuation by cooler particles in the clouds of Venus (Drake, 1966). Cloud attenuation at millimeter wavelengths also gives the best explanation of observed phase variations of brightness temperature (Sagan and Pollack, 1965b). Values of microwave brightness temperature measured over a period of 3 months seemed to correlate with solar activity, but the correlations lacked statistical significance and were judged to be fortuitous (Copeland, 1966).

3.2.4.1.2 Phase Effects: The brightness temperature of Venus is expected to increase as its phase decreases because of the increase of the visible fraction of the illuminated face, which is supposed to be warmer than the dark face. For weak spectral bands the temperature should change very little with phase (Chamberlain, 1965). Several programs of observations to detect phase effects failed because the errors of observation exceeded the small changes in brightness temperature. In some cases, a fictitious reverse

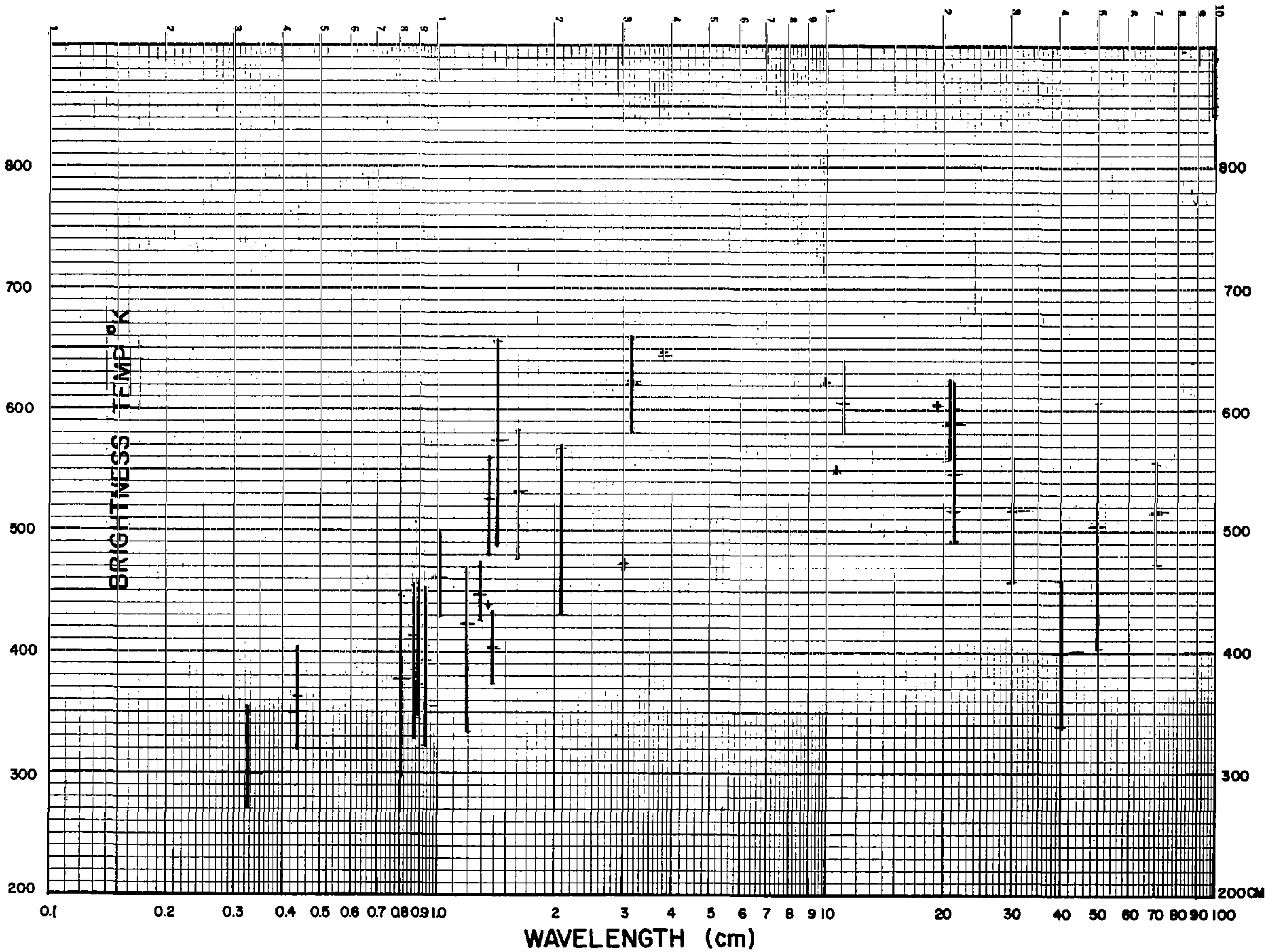

Figure 7. Microwave brightness temperatures with probable errors.

phase effect was noted (Davies and Williams, 1966; Goody, 1965). The microwave brightness temperature at 10.6 cm varied less than 2.5 percent with phase changes (Kuz'min, 1965). The 1.35 cm brightness temperature remained roughly constant when the phase exceeded 130°, but increased at smaller phases to beyond dichotomy (Gibson and Corbett, 1965). Dickel(1966), by means of unconfirmed 3.75 cm microwave measurements, found a sharp rise in blackbody disk temperature as Venus approached superior conjunction (phase of 0°). Theoretically, as well as observationally, the phase changes of brightness temperature are larger at 3 cm than at 10 cm (Pollack and Sagan, 1965a). Results of radio observations of the brightness temperature of the dark and light sides of Venus are summarized as follows (Dickerman, 1966):

TABLE 5. SURFACE TEMPERATURE SUMMARY

Face of Venus	Minimum	Maximum	Most Probable (Owen, 1965)
Dark side	540°K (near pole)	640°K	610° - 640°K
Light side	700°K	800°K	750°K

The unilluminated pole should be colder than the antisolar point because the sun is always low and darkness lasts twice as long at the pole. Clark and Kuz'min(1965) estimated the temperature of the polar regions at 440°K. Plummer and Strong(1966) derived a temperature of only -13°C(260°K) from microwave brightness temperatures corrected for nonthermal radiation.

3.2.4.1.3 Temperature at the Cloud Level: The temperatures shown by infrared radiation are much lower than the surface temperatures determined by microwave radiation because the infrared is being emitted from the perpetual high cloud layer. The most probable temperature of the cloud tops is 235°K (Owen, 1956), which agrees with infrared measurements by Sinton and Strong (1960), but Mariner II data are best fitted by an isothermal cloud layer at 350°K (Jones, 1966). The "hot band" of CO_2 at 8736 to 8768A indicates a temperature of about 400°K, as the emission originates deep within the cloud layer (Spinrad, 1966b). But at the base of the layer, the temperature may be as high as 600°K (Goody and Robinson, 1966) if the clouds have a large vertical extent downward toward the surface. The 3.75 and 8 to 13 micron infrared temperatures are approximately the same on the dark and light sides (Dickerman, 1966): about 240°K at the cloud tops.

3.2.4.2 Stratification. — Limb darkening, in general, indicates a decrease of temperature with increase of height (Goody, 1965). The infrared temperature versus wavelength can be expressed as temperature versus optical depth by use of the infrared absorption spectrum (Chamberlain, 1965). The results show that the cloud tops are cooler than where CO_2 absorption occurs, namely at lower heights. Lapse rates are generally assumed to be adiabatic in the Venusian lower atmosphere (Ho, Kaufman, and Thaddeus, 1966), where thermal equilibrium is more likely to be convective than radiative. In a model where the atmosphere was opaque to solar radiation, a lapse rate of 9K°/km, nearly the adiabatic lapse rate, was chosen (Goody and Robinson, 1966).

If the ionosphere instead of the surface is the source of the microwave radiation of Venus, it would have temperatures as high as 440°K, which if extrapolated downward adiabatically would give a very high surface temperature (Dickerman, 1966). An ionosphere this hot or even hotter, at 600°K (Barrett, 1965), must have a much smaller lapse rate or even a reversal of the vertical temperature gradient at lower altitudes to avoid requiring an unrealistically hot surface.

3.2.4.3 Heat Budget. — The atmosphere of Venus is assumed to be in thermodynamic equilibrium (Lippincott, Eck, Dayhoff, and Sagan, 1966). There are 3 classical models of the atmosphere, summarized by Dickerman(1966): the ionosphere, aeolosphere, and greenhouse models.

The ionosphere model (introduced at the end of the previous section) is the only model which identifies the microwave emission as nonthermal radiation and from a source not at the surface. The solar wind flux at the distance of Venus from the sun is not large enough to supply the ionospheric electron density required to produce the observed intensity of the microwave radiation. Another objection is that Mariner II found no evidence of a strong ionosphere and measured limb darkening instead of the limb brightening required by the model.

In the aeolosphere model, proposed by Öpik(1961), the high surface temperature of Venus is explained by dissipation of the kinetic energy of frequent dust storms through surface friction. The microwave phase effect introduces an inconsistency in the model (Dickerman, 1966). Although the surface does not receive any solar radiation according to this model, brightness temperatures on the illuminated side of Venus are higher than those on the dark side.

The greenhouse model (Sagan, 1960) assumes that some solar radiation can reach the surface and about 99 percent of the surface emission is absorbed by an exceedingly thick atmosphere, almost opaque to infrared radiation (Jastrow, 1966). To have the 1 percent loss to space balance the absorbed solar energy it is necessary to hypothesize a high enough surface temperature to emit a blackbody radiation about 100 times as great. One of the problems of this model is how to account for the great atmospheric opacity. If the surface pressure is about 30 atmospheres, 100 cm of precipitable water and 10^6 cm-atmospheres of carbon dioxide would supply the opacity without the need for clouds, but for smaller values of H_2O and CO_2 contents clouds are necessary to plug up the holes in the principal infrared absorption windows (Ohring and Mariano, 1964 ; Sagan and Pollack, 1966d). Clouds are not only good infrared absorbers, but are also the primary absorbers of microwave radiation (Pollack and Sagan, 1965a). Goody and Robinson (1966) proposed a model atmosphere which is opaque to solar radiation as well as planetary emission. In their model, the cloud tops absorb sunlight on the illuminated side of Venus, become warmer than the clouds on the dark side, and the resulting temperature differences induce a general circulation between the two sides. The related large-scale upward motion accounts for the persistent cloud cover (Goody and Robinson, 1966).

The abundance of water vapor is a key factor in the choice of the most plausible model for the atmosphere of Venus (Drake, 1966). A lack of liquid water on the surface permits high temperatures because there would be no latent heat of vaporization to be lost without evaporation from surface water, such as in an ocean (Dickerman, 1966). The most acceptable model of the Venus atmosphere seems to be the greenhouse model, with a high surface temperature and opacity supplied by clouds (Drake, 1966) consisting mostly of ice crystals.

3.2.5 Pressure and Density

If adiabatic compression plays a part in producing the high temperatures of the lower atmosphere of Venus, one would expect high surface pressures and densities also. There is considerable uncertainty about the atmospheric pressure of Venus. Crude quantitative models lead to values of tens or even hundreds of atmospheres (Ho, Kaufman, and Thaddeus, 1966). From microwave brightness considerations, the surface pressure should lie in the range of 100 to 300 atmospheres if CO_2 is a relatively minor atmospheric constituent and if the temperature lapse rate approximates the adiabatic lapse rate (Ho, Kaufman, and Thaddeus, 1966). At the other extreme, a surface pressure of only 5 atmospheres has been proposed to match a derived low surface temperature (Plummer and Strong, 1965). An intermediate value of surface pressure 10 times this large, namely 50 atmospheres, was chosen by Owen(1965) as most probable within a range of 30 to 100 atmospheres (Dickerman, 1966). If the surface pressure exceeds 60 atmospheres, possibly some small amounts of liquid water may exist on the surface (Dickerman, 1966). The corresponding high densities increase the atmospheric opacity by induced absorption because molecules are sufficiently perturbed by their neighbors to abnegate the usual selection rules (Ho, Kaufman, and Thaddeus, 1966).

Pressure (Dollfus and Maurice, 1965) and particle density (Dollfus, 1966c) are presumed to decrease with increased height according to negative exponential functions. The pressure at cloud-top level is not precisely known. It may be between 0.1 and 1 atmosphere (100 to 1000 mb), perhaps 500 mb, on the illuminated side of Venus, and 90 mb on the dark side, according to Sagan and Kellogg (Dickerman, 1966), but 60 mb has been quoted as the cloud-top pressure on the bright side of Venus (Dollfus, 1966c). The density of the high atmosphere was determined by the differential refraction of the light of Regulas at the occasion of its occultation by Venus in July, 1959. From this density, the pressure at the occultation level of Venus was found to be only 2.5×10^{-3} mb Dickerman, 1966). Altogether, the pressure decreases by 4 or 5 orders of magnitude from the surface up to the occultation level, which shows that the Venusian atmosphere must be deep as well as dense near the surface.

3.3 SURFACE

The surface of Venus is well hidden by its continuous perpetual cloud sheet. Since optical equipment cannot resolve the mysteries below the cloud deck, information must be gained from electromagnetic signals from the surface, which can escape to space. Radar reflections and microwave emission serve the purpose best and will supply the observational information to be presented in the rest of this report.

3.3.1 Polarization

The many potential uses of measurements of polarized thermal radio emission (Drake and Heiles, 1964), described in the section on polarization for Mars, are applicable to the surface of Venus as well. Polarized microwave radiation from Venus was successfully detected and measured (Drake, 1966), even though the upper polarization limit of integral radio emission at a wavelength of 3.75 cm is only 1 percent (Kuz'min and Dent, 1966). Polarization measurements at 10.6 cm during the 1964 inferior conjunction of Venus showed that the emission was from a continuous medium, which was believed to be the surface of the planet (Kuz'min and Klark, 1965). Although the polarization of 10.6 cm radiation from Venus was not appreciable, its distribution over the planet's disk was found to be characteristic of a smooth dielectric sphere, according to Clark and Kuz'min (1965). These measurements have added value when one uses them with the radar-determined dielectric coefficient to estimate what part of the total microwave brightness is due to thermal radiation, which is polarized, and what part is unpolarized nonthermal radiation (Plummer and Strong, 1966). The results have an important bearing on Venus surface temperature determinations.

The reflectivity of a CW radar beam bounced off Venus was found for its two components, the polarized and unpolarized reflectivities, to be 0.114 ± 0.01 and 0.0067 ± 0.0005, respectively (Carpenter, 1966). A normal radar transmits circularly polarized waves and receives the return signal with the opposite sense of polarization, according to Goldstein (1965b). He noted that since a smooth reflector reverses the polarization, the normal mode of radar reception produces the strongest echo response. But multiple reflections are expected where the surface is rough compared to the wavelength. In search of these areas, the same sense of polarization was used for the transmitter and receiver (Goldstein, 1965b) to catch double reversals of polarization.

Optical polarization measurements were made as early as the 1920's by Lyot, who deduced from his observations the diameter of the Venus cloud particles (Dollfus, 1966c). Optical polarization increases with distance from the center of the disk (Gagen-Torn, 1964). To obtain curves of polarization, Dollfus (1966c) made scans from the Venus limb to terminator at several visible and infrared wavelengths. He found that, although the outlines of polarization regions changed continuously and rapidly, there was a tendency for high negative polarization near the supposed poles (Briggs and Mamikunian, 1963). The polarization of Venus was also found to vary with its phase angle, being positive near both inferior and superior conjunctions, with phase angles of 180^{o} and 0^{o} respectively, and most negative near the phase angle of 120^{o} (Dollfus, 1966c). Near inferior conjunction, the directions of optical polarization were nearly parallel to the limb and terminator (Dollfus, 1966c), changing in direction to fit their curvature.

3.3.2 Topography

Radar observations of Venus at a wavelength of 12.5 cm showed that the planet has a mean surface slope of 1/10 (Muhleman, 1966) and is a smooth sphere which, as a whole, deviates very slightly from an isotropic scatterer (Muhleman, 1965). This smooth planet has some rough areas, consisting of mountain ranges or just large fields of boulders, which can depolarize radar signals (Moore, 1966). Multiple radar reflections, indicated by a receiver with the same sense of polarization as the transmitter, are characteristic of a rough surface (Goldstein, 1965b). Anomalies in radar spectrograms, indicative of topographic features, have been used to locate the reflective areas called α and β on the surface of Venus (Goldstein, 1965b). Radar reflections from Venus were stronger in the inferior conjunction of 1964 than in that of 1962, which was explained by assuming that a smoother side of Venus was facing the Earth in 1964 (Kotelnikov, 1965), but this explanation is not valid since the planet presents the same side toward the Earth at successive inferior conjunctions because of the near synchronism of 5 complete Venusian days with this time interval. (See section on period and sense of rotation). Venus is smoother than the Moon, which has had very little erosion, but, with its localized areas of radar scattering enhancement, is not as smooth or as flat as Mars (Pettengill, 1965). Conspicuous topographic prominences are noted by the planetary rotation's doppler effect, which appears as a drift of the corresponding spectral features from the high to low frequency side of the spectrum (Goldstein, 1966b). The rate of frequency drift yields the Venusian latitude of the topographic feature (Goldstein, 1966b) and the timing of the frequency changes gives the longitude. The only remaining ambiguity is whether the prominence is in the N or S hemisphere of Venus. This doubt can be resolved by measuring the length of time required for the feature to travel from limb to limb provided the subpoint of the observer is not on the Venusian equator.

3.3.3 Composition of the Surface

The dielectric constant of the surface of a planet, measured from microwave radiation, is an indicator of the composition of the surface materials. There is considerable doubt about the accuracy of the measurements of the dielectric constant and conductivity of the Venusian surface, the derived dielectric constant ranging between 2.2 and 7.1 (Francis, 1965). A derived range of the dielectric constant between 3 and 7 indicates that there are no large bodies of water on Venus (Dickerman, 1966). A surface dielectric constant between 4 and 5 not only rules out significant amounts of liquid water, but is consistent with a surface of dry rocks (Pettengill, 1965). Estimates of the dielectric constant of Venus with smaller probable error were 3.75 ± 0.3, obtained by Carpenter (1966), and 2.2 ± 0.2, by Barrett (1965). When corrected for the roughness of the surface, the latter value is raised to 2.5 (Clark and Kuz'min, 1965), the same dielectric constant as that given by Drake (1966), who interpreted it as an indication of loose material on the surface. These calculated dielectric constants have been criticized by Plummer and Strong (1966) as being too low because of the supposed admixture of nonthermal radiation with the surface thermal radiation at microwave frequencies. They surmised lower values of the latter and consequently a low enough surface temperature to permit high-latitude snowfall of sufficient amount to produce ice caps, for which there is no observational evidence.

Dry terrestrial soils have dielectric constants similar to the Venusian values, which do not differ much from those for the Moon, Mercury, and Mars (Ho, Kaufman, and Thaddeus, 1966). There are 2 conflicting theories about the nature of the surface of Venus: a sterile, dusty desert with no trace of moisture anywhere or a surface covered almost entirely with water (Moore, 1966). The microwave observations showed a high surface temperature, supporting the dry desert theory. Any existing areas with liquids are probably small and may be turbulent oceans of hydrocarbons (Dickerman, 1966). A comparison of the Venusian spectrum with laboratory data, however, shows that an abundance of hydrocarbons is excluded along with magnetic materials and granite, whereas the primary surface materials may be fused quartz and a wide range of powdered oxides, carbonates, and silicates (Pollack and Sagan, 1965a). By analogy with the Earth, iron should be more abundant than sulfur in the crust of Venus, and this should result in iron compounds controlling the degree of oxidation as well as the abundances of sulfur compounds (Mueller, 1965). A rather startling view of materials of a hot surface of Venus is presented by Owen (1965) as follows. Since the melting points of tin, lead, aluminum, magnesium, bismuth, and zinc might be reached on the sunlit side, molten pools of these metals could cover much of that area. With the high surface pressures, some constituents of the lower atmosphere may condense out. Therefore, seas around the dark pools may contain benzine, acetic acid, butyric acid, and phenol in the liquid state (Owen, 1965). The surface of Venus may not be quite as forbidding as this if the abundances of the compounds are very low, but nevertheless the high temperature is a very undesirable property for manned exploration.

4. GENERAL OUTLOOK

According to Menzel (1966), who has read hundreds of papers and a number of books on the planets, "it is indeed an understatement to say that the disagreement among various workers is fundamental and widespread." Perhaps some day this will become an overstatement, as more differences are resolved. Approaches on diverse lines of attack should continue so that there will be more theories from which to select the best choices. Observational progress will get a big boost from on-the-spot measurements within the atmospheres or on the surfaces of the planets. The future looks favorable also on the basis of accomplishments so far this decade. No wonder the outlook is so optimistic!

REFERENCES

Adamcik, J. A., 1963: The water vapor content of the Martian atmosphere as a problem of chemical equilibrium. Planetary and Space Science, v. 11, 355-359.

Adelman, B., 1966: Altitude variations on Mars. The Rev. of Popular Astronomy, v. LX, no. 540, p. 20.

Anders, E., and J. R. Arnold, 1965: Age of craters on Mars. Science, 149, 1494-1496.

Anderson, A. D., 1965: A model for the lower atmosphere of Mars based on Mariner IV occultation data. Lockheed Palo Alto Res. Lab. 6-75-65-62, 26 pp.

__________, 1966: Spherical particle terminal velocities in the Martian daytime atmosphere from 0 to 50 km. Lockheed Missiles and Space Co., Palo Alto, Cal., LMSC-6-76-66-21, 37 pp.

Applebaum, D. C., P. Harteck, R. R. Reeves, Jr., and B. A. Thompson, 1966: Some comments on the Venus temperature. J. Geophys. Res., 71, 5541-5545.

Barabashov, N. P., 1965: Optical properties of the atmosphere of Mars based on photometric observation data. In: Akademiia Nauk Ukrainskoĭ RSR, 52-60.

Barath, F.T., A. H. Barrett, J. Copeland, D.E. Jones, and A.E. Lilley, 1964: Mariner 2 microwave radiometer experiment and results. Astronomical J., 69, 49-58.

Barrett, A.H., 1965: Passive radio observations of Mercury, Venus, Mars, Saturn, and Uranus. Radio Science, Nat. Bureau of Standards, 1565-1573.

Basharinov, A.E., and B.G. Kutuza, 1965: Nature of the cloud layer of Venus- From radiometric observations in the SHF Range. Astronomicheskii Zhurnal, 43, 149-153.

Belton, M.J.S., and D. M. Hunten, 1966a: The abundance and temperature of CO_2 in the Martian atmosphere. Astrophys. J., 145, 454-467.

__________, 1966b: Water vapor in the atmosphere of Venus. Astrophys. J., 146, 307-308.

Blum, R., 1965: Magnetosphere and the Martian blue clearing. Nature, 207, 1343-1344.

Bottema, M., W. Plummer, and J. Strong, 1964: Water vapor in the atmosphere of Venus. Astrophys. J., 139, 1021-1022.

REFERENCES (Cont.)

Bottema, M., W. Plummer, J. Strong, and R. Zander, 1964: Composition of the clouds of Venus. Astrophys. J., 140, p. 1640.

Briggs, M. H., and G. Mamikunian, 1963: Venus: A summary of present knowledge. Jet Propulsion Lab. under NASA Contract NAS7-100, NASA CR-50873, 39 pp.

Cann, M.W.P., W. O. Davies, J.A. Greenspan, and T.C. Owen, 1965: A review of recent determinations of the composition and surface pressure of the atmosphere of Mars. IIT Res. Inst. under Contract NAS5-9037, NASA CR-298.

Carpenter, R.L., 1966: Study of Venus by CW radar - 1964 results. Astronom. J., 71, 142-152.

Chamberlain, J. W., 1965: Atmosphere of Venus near her cloud tops. Astrophys. J., 141, 1184-1205.

Clark, B.G., and A.D. Kuz'min, 1965: Measurement of the polarization and brightness distribution of Venus at 10.6-cm wavelength. Astrophys. J., 142, 23-44.

Copeland, J., 1966: Observations of Venus at 8.6 mm. λ near inferior conjunction. Astrophys. J., 143, 996-1000.

Davies, R.D., and D. Williams, 1966: Observations of the continuum emission from Venus, Mars, Jupiter, and Saturn at 21.2-cm wavelength. Planetary and Space Sci., 14, 15-32.

Deirmendjian, D., 1966: Comments on the detection of water and ice clouds on Venus. In Proc. of the Caltech-JPL Lunar and Planetary Conference, Sept. 13-18, 1965, 150-154.

de Vaucouleurs, G., 1961: Reconnaissance of the nearer Planets. Air Force Office of Scientific Res. AFOSR/DRA-61-1.

Dickel, J.R., 1966: Measurement of the temperature of Venus at a λ of 3.75 cm for a full cycle of planetary phase angles. Icarus, 5, 305-308.

Dickerman, P., 1966: Scientific objectives of deep space investigations - Venus. Report No. P-7, IIT Res. Inst.

Dollfus, A., and E. Maurice, 1965: Study of the elongation of the crescent cusps of Venus in June 1964. Académie des Sciences, Comptes Rendus, 260, 427-430.

Dollfus, A., 1965: Study of the planet Mars, 1954-1958. Annales d'Astrophysique, 28, 722-747.

REFERENCES (Cont.)

Dollfus, A., 1966a: Determination of the luminance of the Martian atmosphere. Académie des Sciences, Comptes Rendus, 262, 519-522.

__________, 1966b: Contribution au Colloque Caltech-JPL sur la lune et les planètes: Mars. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 288-305.

__________, 1966c: Contribution au Colloque Caltech-JPL sur la lune et les planètes: Venus. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 187-202.

Dollfus, A., and J. Focas, 1966a: Purity of the Martian atmosphere. Académie des Sciences, Comptes Rendus, 262, 1024-1027.

__________, 1966b: Polarimetric study of the planet Mars. Observatoire de Paris, Final Scientific Report, Contract AF 61(052)-508.

Donahue, T. M., 1966: Upper atmosphere and ionosphere of Mars. Science, 152 763-764.

Drake, F.D., and C.E. Heiles, 1964: Analysis of planetary surface parameters through observation of polarized thermal radio emission. Astronomical J., 69, p. 539.

Drake, F.D., 1966: Recent radio observations of the planet Venus. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 180-183.

Draper, A.L., and J.A. Adamcik, 1966: Spectral studies of materials possibly present on the Martian surface. Texas Tech. College, Final Rpt. No. 1, 11 pp.

Dyce, R.B., 1965: Recent Arecibo observations of Mars and Jupiter. In Radio Science, 1628-1629.

__________, 1966: Some radar characteristics of Venus at 430 MHz. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 172-177.

Epstein, E.E., 1966: Disk temperatures of Mercury and Mars at 3.4 mm. Astrophys. J., 143, 597-598.

Eshleman, V.R., 1966: The Stanford study of the Mariner IV occultation data. Semiannual Report No. 3, 3 pp.

Evans, D.C., 1965: Ultraviolet reflectivity of Mars. Science, 149, 969-972.

__________, 1966: UV reflectivity of Venus and Jupiter. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966, 28 pp.

REFERENCES (Cont.)

Fjeldbo, G., W. C. Fjeldbo, and V. R. Eshleman, 1966a: Models for the atmosphere of Mars based on the Mariner IV occultation experiment. Scientific Reports Nos. 2 and 15, 36 pp.

__________, 1966b: Atmosphere of Mars: Mariner IV models compared. Science, 153, 1518-1523.

__________, 1966c: The atmosphere of Mars: A comparison of different model studies based on Mariner IV occultation data. Scientific Rpts. Nos. 3 and 16, 24 pp.

Fjeldbo, G., V.R. Eshleman, A.J. Kliore, D.L. Cain, G.S. Levy, and F.D. Drake, 1966: Preliminary results of the Mariner IV radio occultation measurements of the upper atmosphere of Mars. In Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 267-272.

Francis, C.E., 1965: The electromagnetic propagation characteristics of Venus and Mars. IEEE Proc., v. 53, 1216-1221.

Gagen-Torn, V.A., 1964: Photographic polarimetry of Venus. Leningradskii Gosudarstvennyi Universitat, Trudy, v. 21, 95-98.

Gibson, J.E., and H.H. Corbett, 1965: Radiation of Venus at the 13.5-mm water-vapor line. In Radio Science, J. of Res., Nat. Bureau of Standards, 1577-1579.

Gifford, F.A., Jr., 1964: A study of Martian yellow clouds that display movement. Monthly Weather Rev., 92, 435-440.

Goldreich, P., and S.J. Peale, 1966: Resonant rotation for Venus? Nature, 209, 1117-1118.

Goldstein, R.M., and W.F. Gillmore, 1963: Radar observations of Mars. Science, 141, p. 1171.

Goldstein, R.M., 1965a: Mars - Radar observations. Science, 150, 1715-1717.

__________, 1965b: Preliminary Venus radar results. In Radio Science, J. of Res., Nat. Bureau of Standards, 1623-1625.

__________, 1966a: JPL radar observations of Mars. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, p. 239.

__________, 1966b: Radar studies of Venus. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966, 12 pp.

Goody, R., 1965: The structure of the Venus cloud veil. J. Geophys. Res., 70, 5471-5482.

REFERENCES (Cont.)

Goody, R.M., and A.R. Robinson, 1966: A discussion of the deep circulation of the atmosphere of Venus. Astrophys. J., 146, 339-355.

Goy, G., 1966: The Planet Mars. Sciences et Industries Spatiales, 2, 69-74.

Gray, L.D., 1966a: Interpretation of low resolution spectra of Mars in the 2μ region. AIAA Aerospace Sciences Meeting, New York, Jan. 24-26, 1966, 12 pp.

__________, 1966b: Transmission of the atmosphere of Mars in the region of 2μ. Icarus, 5, 390-398.

Greenspan, J.A., 1965: Mars: compatible determinations of surface pressure through particle scattering. Science, 150, 1156-1158.

Gross, S.H., W.E. McGovern, and S.I. Rasool, 1966: Mars - Upper atmosphere. Science, 151, 1216-1221.

Hibben, R.D., 1965: Some Earth life seen adaptable to Mars. Aviation Week and Space Tech., 82, 71-79.

Ho, W., I.A. Kaufman, and P. Thaddeus, 1966: Laboratory measurement of microwave absorption in models of the atmosphere of Venus. J. Geophys. Res., 71, 5091-5108.

Horowitz, N.H., 1966: The search for extraterrestrial life. Science, 151, 789-792.

Hunten, D.M., 1966: CO_2 bands and Martian surface pressure. In Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 240-244.

Ianovitskiĭ, E.G., 1965: Aerosol component of the Martian atmosphere. In Akademiia Nauk Ukrainskoĭ RSR, 82-90.

Jastrow, R., 1966: The planet Venus. Science J., 2, 58-62.

Johnson, F.S., 1965: Atmosphere of Mars. Science, 150, 1445-1448.

Jones, D.E., 1966: The Mariner II microwave radiometer experiment. Jet Propulsion Lab., NASA Tech. Rpt. No. 32-722.

Kachur, V., 1965: Anomalous influence of atmosphere in thermology of the Martian surface. J. Geophys. Res., Letters Section, Oct. 17, 1965.

__________, 1966a: Thermological aspects of the Martian surface environment. In Proc. of IES Annual Tech. Meeting, San Diego, April 13-15, 1966.

Kachur, V., 1966b: Thermology of the Martian surface. AIAA Aerospace Sciences Meeting, Los Angeles, June 27-29, 1966, 8 pp.

Kaplan, L.D., G. Münch, and H. Spinrad, 1964: An analysis of the spectrum of Mars. Astrophys. J., 139, 1-15.

Kaplan, L.D., 1964: Venus, recent physical data for June 1964. Jet Propulsion Lab., CIT, NASA-CR-56851, 6 pp.

Katterfel'd, G.N., 1965: Volcanic activity on Mars? Priroda, Moscow, 8, 103-109.

Kellermann, K.I., 1965: Radio observations of Mercury, Venus, Mars, Saturn, and Uranus. In Radio Science, J. of Res., Nat. Bureau of Standards, 1574-1575.

Kiess, C.C., S. Karrer, and H.K. Kiess, 1960: A new interpretation of Martian phenomena. Publications of the Astronomical Soc. of Pacific, 72, 256-267.

Kliore, A., D.L. Cain, G.S. Levy, V.R. Eshleman, G. Fjeldbo, F.D. Drake, 1965: Occultation experiment: Results of the first direct measurement of Mars' atmosphere and ionosphere. Science, 149, 1243-1248.

__________, 1966: Preliminary results of the Mariner IV occulation measurement of the atmosphere of Mars. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 257-266.

Kliore, A., D.L. Cain, and G.S. Levy, 1966: Radio occultation measurement of the Martian atmosphere over 2 regions by the Mariner IV space probe. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966, 18 pp.

Kotelnikov, V.A., 1965: Radar observations of Venus in the Soviet Union in 1964. In Radio Science, J. of Res., Nat. Bureau of Standards, 1634-1636.

Kozyrev, N.A., 1964: Spectral evidence of the existence of snow in the Martian atmosphere. Life Sciences and Space Res., 2, Amsterdam, p.250.

Kuiper, G.P., 1964: Infrared spectra of stars and planets, IV: The spectrum of Mars, 1-2.5μ, and the structure of its atmosphere. Communications of the Lunar and Planetary Lab., 2, p. 79.

Kutuza, B.G., B. Ia. Losovskii, and A.E. Salomonovich, 1966: Observation of Martian radio emission at 8 mm. Astronomicheskii Zhurnal, 43, 236-237.

Kuz'min, A.D., and B. D. Klark, 1965: Measurements of polarization and of brightness temperature distribution of Venus in the 10.6 cm wave. Akad. Nauk SSSR, Astronomiya, 161, 551-553.

Kuz'min, A.D., 1965: Measurement of the brightness temperature of the illuminated side of Venus at 10.6 cm. Astronomicheskiĭ Zhurnal, Moscow, 42, 1281-1286.

Kuz'min, A.D., and W.A. Dent, 1966: Measurements of the brightness temperature and polarization for Venus' radio emission at a λ of 3.75 cm. Astronomicheskiĭ Zhurnal, 43, 692-694.

Kuz'min, A.D., 1966: Some remarks concerning the radioastronomical observations of Venus. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 184-186.

Leighton, R.B., 1966: Comments on the Mariner IV photographs of Mars. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18,1965, 273-278.

Leighton, R.B., and B.C. Murray, 1966: Behavior of CO_2 and other volatiles on Mars. Science, 153, 136-144.

Leovy, C., 1966a: Note on thermal properties of Mars. Icarus, 5, 1-6.

__________, 1966b: Radiative-convective equilibrium calculations for a two-layer Mars atmosphere. RAND Corp. under Contract NAS-21(07), 55 pp.

Lippincott, E.R., R.V. Eck, M.O. Dayhoff, and C. Sagan, 1966: Thermodynamic equilibria in planetary atmospheres. In: National Biomedical Res. Found., Invest. of thermodyn. mech. for the production of complex compds. essent. for the origin of life, 34 pp.

Loomis, A.A., 1965: Some geologic problems of Mars. Geological Soc. of Am. Bulletin, 76, 1083-1103.

Marmo, F.F., Shardanand, and P. Warneck, 1965: Ozone distribution in the atmosphere of Mars. J. Geophys. Res., 70, 2270-2272.

McElroy, M.B., J. L'Ecuyer, and J.W. Chamberlain, 1965: Structure of the Martian upper atmosphere. Astrophys. J., 141, 1523-1535.

McLaughlin, D.B., 1956: New interpretation of the surface of Mars. Sci. Monthly, 83, p. 176.

Menzel, D.H., 1963: The atmosphere of Mars. Phys. of the Planets, 11th Internat. Astrophys. Symp., Liege, July 9-12, 1962, 411-414.

Menzel, D.H., 1966: The surfaces of the Moon, Mars and Venus. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966.

Mintz, Y., 1961a: The general circulation of planetary atmospheres. In The Atmospheres of Mars and Venus. Space Sci. Board, Nat. Acad. Sci., 107-146.

__________, 1961b: Temperature and circulation of the Venus atmosphere. Planetary and Space Sci., 5, 141-152.

Miyamoto, S., 1963: Observational study on the general circulation of Mars. U. of Kyoto, Inst. of Astrophys. and Kwasan Obs. Contrib. No. 125, 81-88.

__________, 1965: Meteorological observations of Mars during the 1965 opposition. U. of Kyoto, Inst. of Astrophys. and Kwasan Obs. Contrib. No. 141, 45pp.

__________, 1966: Martian atmosphere and crust. Icarus, 5, 360-374.

Moore, P., 1966: Mysteries of Venus. Rev. of Popular Astronomy, LX, 8-10.

Moore, R.C., and G.F. Schilling, 1965: On the scattering of sunlight into planetary shadow cones. Presented at Symp. on Interdisciplinary Aspects of Radiative Energy Transfer, Philadelphia, 24-26 February 1966.

Moroz, V.I., 1964: The infrared spectrum of Mars ($\lambda 1.1 - 4.1\mu$). Astronomicheskiĭ Zhurnal, Moscow, 41, 350-361.

Morozhenko, A.V., and E.G. Ianovitskiĭ, 1965: Determination of the optical parameters of the Martian atmosphere and surface taking into account anisotropic scattering. In Problems of Astrophysics, ed. I.K. Koval', Izdatel'stvo Naukova Dumka, 127-131.

Mueller, R.F., 1964: A chemical model for the lower atmosphere of Venus. Icarus, 3, 285-298.

__________, 1965: Stability of sulfur compounds on Venus. Icarus, 4, 506-512.

Muhleman, D.O., 1965: Radar scattering from Venus and Mercury at 12.5 cm. In Radio Science, J. of Res., Nat. Bureau of Standards, 1630-1631.

__________, 1966: Venus radar investigations. In Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 164-171.

Myer, H.G., J. T. Ohrenberger, and T. R. Thompson, 1965: Emission and absorption of radiant energy in a model planetary atmosphere. AIAA J., 3, 2203-2210.

Neubauer, F.M., 1966: Thermal convection in the Martian atmosphere, J. of Geophys. Res., 71, 2419-2426.

Nicolson, I., and P. Moore, 1964: The phase anomaly of Venus. J. of Brit. Astronomical Assoc., 74, 269-271.

Ohring, G., and J. Mariano, 1964: The effect of cloudiness on a greenhouse model of the Venus atmosphere. J. of Geophys. Res., 69, 165-175.

Ohring, G., W. Tang, and J. Mariano, 1965: Planetary meteorology. GCA Corp., Contract NASw-975, GCA Tech. Rpt. No. 65-9-N, 88 pp.

Ohring, G., and J. Mariano, 1966: The vertical temperature distribution in the Martian atmosphere. J. of Atmos. Sci., 23, 251-255.

Ohring, G., W. Tang, F.B. House, and J. Mariano, 1966: Contributions to planetary meteorology, GCA Final Report, GCA Tech. Rpt. No. 66-8-N, 87 pp.

Ohring, G., 1966: Water-vapor mixing ratios near the cloudtops of Venus. Icarus, 5, 329-333.

O'Leary, B., 1965: Revised upper limit of NO_2 in the Martian atmosphere. Astronomical Soc. of Pacific, Pub. 77, 167-177.

__________, 1966: The presence of ice in the Venus atmosphere as inferred from a halo effect. Astrophys. J., 146, 754-766.

Öpik, E.J., 1961: The aeolosphere and atmosphere of Venus. J. of Geophys.Res., 66, 2807-2819.

__________, 1963: Venus and the Mariner. Irish Astronomical J., 6, 59-67.

__________, 1966: The Martian surface. Science, 153, 255-265.

Otterman, J., and F.E. Bronner, 1966: Martian wave of darkening: A frost phenomenon? Science, 153, 56-60.

Owen, R.B., 1965: Theoretical model atmospheres of Venus, NASA Tech. Note D-2527, 41 pp.

Owen, T., 1966a: Recent observations of the photographic spectrum of Mars - A preliminary result. In Proc. of Caltech-JPL Lunar and Planetary Conf. Sept. 13-18, 1965, 247-251.

Owen, T., 1966b: The composition and surface pressure of the Martian atmosphere: Results from the 1965 Opposition. <u>Astrophys. J.</u>, 146, 257-270.

Paetzold, H.K, 1963: On the problems of the Martian ozonosphere. In Société Royale des Sciences de Liège, <u>Phys. of the Planets</u>, 11th Internat. Astrophys. Symp., Liège, July 9-12, 1962, 452-459.

Palm, A., and B. Basu, 1965: Blue haze of Mars. <u>Icarus</u>, 4, 111-118.

Pearce, T., 1964: The Venus cusp caps as aurora - an objection. <u>J. of Brit. Astronomical Assoc.</u>, 74, 301-302.

Pettengill, G.H., 1965: A review of radar studies of planetary surfaces. In <u>Radio Science</u>, J. of Res., Nat. Bureau of Standards, 1617-1623.

Pettit, E., 1961: The solar system, Vol. 3, <u>Planets and Satellites</u>, ed. G.P. Kuiper and B.M. Middlehurst, U. of Chic. Press, Chap. X.

Pither, C.M., 1962: The cusp caps of Venus. <u>J. of Brit. Astronomical Assoc.</u>, 72, 14-17.

__________, 1963: The origin of the Cytherean cusp caps. <u>J. of Brit. Astronomical Assoc.</u>, 73, 197-199.

Plummer, W.T., and J. Strong, 1965: Conditions on the planet Venus. <u>Astronautica Acta</u>, 11, 375-382.

__________, 1966: A new estimate of the surface temperatures of Venus. <u>Astrophys. J.</u>, 144, 422-423.

Pollack, J.B., and C. Sagan, 1965a: The microwave phase effect of Venus. <u>Icarus</u>, 4, 62-103.

__________, 1965b: Infrared limb darkening of Venus. <u>J. of Geophys. Res.</u>, 70, 4403-4426.

__________, 1966: Secular changes and dark-area regeneration on Mars. Smithsonian Inst. Astrophys. Obs. 7-66-31.

Pollack, J.B., E.H. Greenberg, and C. Sagan, 1966: A statistical analysis of the Martian wave of darkening and related phenomena. Smithsonian Inst. Astrophys. Obs. 8-66-20.

Prabhakara, C., and J.S. Hogan, Jr., 1965: Ozone and carbon dioxide heating in the Martian atmosphere. <u>J. of Atmos. Sci.</u>, 22, 97-109.

REFERENCES (Cont.)

Prokof'yev, V.K., and N.N. Petrova, 1964: On the presence of oxygen in the atmosphere of Venus. Jet Propulsion Lab, Contract NAS7-100, NASA CR-53061, 21 pp.

Prokof'yev, V.K., 1965: Spectroscopic studies of the composition of the Cytherean atmosphere. In Problems of Astrophysics, ed. I.K. Koval', 6-11.

Rakos, K.D., 1965: The atmospheric Pressure at the surface of Mars. Lowell Obs., NASA-CR-71312, Flagstaff, Ariz., 15 pp.

Rea, D.G., 1965: The atmosphere and surface of Mars. A selective review. Presented at Lunar and Planetary Seminar, C.I.T., Sept. 17, 1965.

Robinson, J.C., 1966: Ground-based photography of the Mariner IV region of Mars. Icarus, 5, 245-247.

Ryan, J.A., 1964: Notes on the Martian yellow clouds. J. Geophys. Res., 69, 3759-3770.

Sagan, C., 1960: Radiation balance of Venus. CIT, Jet Propulsion Lab., Contract NASw-6, Tech. Rpt. 32-34, 23 pp.

__________, 1965: Origins of the atmospheres of the Earth and planets. Section I. Origin of the Earth, of the International Dictionary of Geophysics, S.K. Runcorn, London Pergamon Press.

Sagan, C., and J.B. Pollack, 1965a: Spacecraft observation of Venus infra-red limb darkening. Annales d'Astrophysique, 28, 229-233.

__________, 1965b: An analysis of microwave observations of Venus. In Radio Science, J. of Res., Nat. Bureau of Standards, 1583-1584

Sagan, C., J.P. Phaneuf, and M. Ihnat, 1965: Total reflection spectro-photometry and thermogravimetric analysis of simulated Martian surface materials. Icarus, 4, 43-61.

Sagan, C., 1966a: Mariner IV observations and the possibility of iron oxides on the Martian surface, 1966. Icarus, 5, 102-103.

__________, 1966b: On the nature of the canals of Mars. Smithsonian Inst. Astrophys. Obs. 8-66-17.

__________, 1966c: The Mariner IV mission to Mars. Astronomical Soc. of Pacific Leaflet, 8 pp.

Sagan, C., and J.B. Pollack, 1966a: In inorganic model of Martian phenomena. Astrophys. J., 71, no. 1338.

REFERENCES (Cont.)

Sagan, C., and J.B. Pollack, 1966b: Radio and radar evidence on the structure and composition of the Martian surface. In: Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 255-256.

__________, 1966c: Properties of the clouds of Venus. In: Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965.

__________, 1966d: On the nature of the clouds and the origin of the surface temperature of Venus. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966.

__________, 1966e: Anisotropic non-conservative scattering and the clouds of Venus. Smithsonian Inst. Astrophys. Obs. 8-66-22.

__________, 1966f: Elevation differences on Mars. Smithsonian Inst. Astrophys. Obs.

Sagan, C., J.B. Pollack, and R.M. Goldstein, 1966: Radar Doppler spectroscopy of Mars. 1. Elevation differences between bright and dark areas. Res. in space science, Smithsonian Inst. Astrophys. Obs., Spec. Rpt. No. 221.

Sharanov, V.V., 1965: Some remarks on the hypothesis of an absorbing atmosphere of Mars. In Akad. Nauk Ukrainskoĭ RSR, 61-71.

Shawhan, S.D., 1966: Ionization rate and electron density profile for the Martian ionosphere based on Mariner IV observations. 47th Annual AGU Meeting, April 19-22, 1966, Wash., D.C.

Shimizu, M., 1966: A theoretical investigation of Cytherian, Martian, and Jovian upper atmospheres. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966, 4 pp.

Simpson, J.F., 1966a: Evidence for the volcanic origin of lunar and Martian craters. Earth and Planetary Science Letters, Vol. 1, 130-131.

__________, 1966b: Additional evidence for the volcanic origin of lunar and Martian craters. Earth and Planetary Science Letters, Vol.1, 132-134.

Sinton, W.M., and J. Strong, 1960: Radiometric observations of Venus. Astrophys. J., 131, 470-490.

Sinton, W.M., 1961: The solar system, Vol. 3, Planets and Satellites, ed. G.P. Kuiper and B.M. Middlehurst, U. of Chic. Press, Chap. XI.

__________, 1963: Infrared observations of Venus. Proc. of 11th Internat. Astrophys. Symp., Liège, July 9-12, 1962, 300-310.

Sinton, W.M., 1964: Physical researches on brighter planets. Lowell Obs., Contract No. AF 19(604)-5874, Flagstaff, Ariz., Final Report.

Sloan, R.K., 1966: The scientific experiments of Mariner IV, Scientific American, 214, 2-12.

Smith, B.A., 1963: Photographic appearance of Venus at the time of Mariner II encounter. J. Geophys. Res., 68, 4363-4365.

Smith, B.A., and W.E. Bains, 1964: Observations of the optically thin scattering layer in the atmosphere of Venus. Astronomical J., 69, p. 150.

Smoluchowski, R., 1965: Is there vegetation on Mars? Science, 148, 946-947.

Spencer, D.F., 1966: Our present knowledge of the Martian atmosphere. In Proc. of AIAA-AAS Stepping Stones to Mars Meeting, Baltimore, March 28-30, 1966.

Spinrad, H., 1963: Spectroscopic temperature and pressure measurements in the Venus atmosphere. Pub. of Astronomical Soc. of Pacific, Vol. 74, 187-201.

Spinrad, H., and E.H. Richardson, 1965: An upper limit to the molecular oxygen content of the Venus atmosphere. Astrophys. J., 141, 282-286.

Spinrad, H., 1966a: Spectroscopic Observations of Mars. In Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 245-246.

__________, 1966b: Resolution of a CO_2 "Hot Band" in the Venus spectrum. Astrophys. J., 145, 943-945.

Spinrad, H., and S.J. Shawl, 1966: Water vapor on Venus - a confirmation. Astrophys. J., 146, 328-329.

Spinrad, H., R.A. Schorn, R. Moore, L.P. Giver, and H.J. Smith, 1966: High-dispersion spectroscopic observations of Mars. I. The CO_2 content and surface pressure. The Astrophys. J., 146, 331-338.

Strong, J., 1966: Balloon telescope studies of Venus. In Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 147-148.

Strughold, H., 1966: A new look at Mars - Medico-biological aspects. TRW Space Log, Winter 1965-66, 2-13.

Suess, H.A., 1963: Remarks concerning the chemical composition of the atmosphere of Venus. U. of Calif., NASA-CR-55367, 12 pp.

REFERENCES (Cont.)

Tang, W., 1965: Some aspects of the atmospheric circulation on Mars. GCA Tech. Rpt. No. 65-4-N, 43 pp.

__________, 1966: A study of the general circulation of the Martian atmosphere based upon the result of the occultation experiment from Mariner IV. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966.

Tepper, M., 1965: The systematic investigation of the meteorology of Mars. In Proc. of Conf. on the Exploration of Mars and Venus, Blacksburg, Va., Aug. 23-27, 1965.

Thomson, J.H., 1966: Radar observations of Venus at Jodrell Bank. In Proc. of Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965, 178-179.

Tolbert, C.W., 1966: Observed millimeter wavelength brightness temperatures of Mars, Jupiter, and Saturn. Astronomical J., 71, 30-32.

Van Tassel, R.A., and J. W. Salisbury, 1965: Planetary environments. In: Handbook of Geophysics and Space Environments, AFCRL, Chap. 20, 19 pp.

Weaver, H., 1965: Some problems of planetary radio astronomy. In: NATO Advanced Study Institute of the Nat. Obs. of Athens, Plenum Press, 371-399.

Weertman, J., 1966: The effect of a low viscosity layer on convection in the mantle. Army Cold Regions Res. and Engn. Lab., CRREL-RR-203, 26 pp.

Weil, N.A., 1965: Lunar and planetary surface conditions. Academic Press, 222 pp.

Wells, R.A., 1965a: Re-evaluation of W.H. Wright's plates of the 1924 and 1926 oppositions of Mars. Planet. and Space Sci., 13, 261-263.

__________, 1965b: Evidence that the dark areas on Mars are elevated mountain ranges. Nature, 207, 735-736.

__________, 1966: An analysis of Martian clouds and their topographical relationship. COSPAR, Internat. Space Science Symposium, Vienna, May 10-19, 1966, 59 pp.

Younkin, R.L., 1966: A search for Limonite near-IR spectral features on Mars. In Proc. of the Caltech-JPL Lunar and Planetary Conf., Sept. 13-18, 1965. p. 252.

NASA-Langley, 1967 —— 30 CR-786