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SOME ASPECTS OF THE ATMOSPHERIC CIRCULATION ON MARS

by W. Tang

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GCA CORPORATION
Bedford, Mass.
for

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SOME ASPECTS OF THE ATMOSPHERIC CIRCULATION ON MARS

By Wen Tang

SUMMARY

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Inferences concerning atmospheric circulation features on Mars are made from analysis and interpretation of some observed Martian cloud systems and from application of meteorological theory to the Martian atmosphere. The trajectories of several cloud systems and the use of two different theoretical criteria suggest the presence of a wave type circulation regime in the mean for the year on Mars. Cloud observations also suggest the presence of sub-tropical high pressure centers, upper level meridional flow at equatorial latitudes, and frontal cloud phenomena at equatorial latitudes. Theoretical estimates of the mean large scale zonal and meridional wind velocities yield values greater than on earth. Theoretical estimates of the maximum surface wind suggest a value greater than 100 m sec^{-1}. Computations of the vertical velocity profile, based upon a simplified "ω" equation, indicate greater large scale vertical velocities than on earth, and a "dynamic" tropopause height of about 20 km.

Author

INTRODUCTION

Among the planets in the solar system, Mars is probably the one whose meteorology resembles that of the earth most closely. This is mainly due to the fact that the basic physical parameters on Mars, such as its size, axial inclination, angular velocity and distance from the sun, are not too much different from earth's. Therefore, based on the same meteorological principles and theories for earth, one may infer some aspects of Martian meteorology. The relative closeness of the planet to the earth permits astronomical observations that are useful in the development and testing of theoretical models.

Based on some radiometric and cloud movement observations, Hess [1]* constructed the first surface temperature and flow pattern maps for Mars. The patterns on his isothermal and stream line charts resembled patterns in the terrestrial atmosphere. Mintz [2] made use of a theoretical model to study the circulation of the Martian atmosphere. Studying the dynamic stability of a symmetrical regime, he concluded that, in the mean for year on Mars, the general circulation will be in the symmetrical regime and dynamically stable.

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*Numbers in [ ] throughout the text indicate reference numbers.

The purpose of the present work is to utilize some recent data as well as past observations to study various problems related to the circulation of the Martian atmosphere. These problems include the magnitude of the mean zonal wind velocity, the possible maximum surface wind speed in storms, the profile of the vertical velocity, and the estimated height of the tropopause.

In Section 1, some inferences are made from cloud observations about storm tracks, position of subtropical high center, and the estimated height of the tropopause. The slope of a front in the Martian tropical region is computed and is discussed in Section 2. In Section 3, the mean annual circulation regime is predicted from estimates of the magnitude of the thermal Rossby number and the rotation parameter. In Section 4, the mean and maximum wind velocities in a Martian storm are computed, based on recent data on the surface pressure and the radiation budget. In Section 5, an estimate of the vertical velocity profile is presented. This estimate is based upon computations with an adiabatic model in which a mean zonal velocity profile is assumed. From the vertical velocity profile, estimation of the height of the tropopause on Mars is derived.

1. INFERENCES FROM CLOUDS

Knowledge of clouds is of great value for understanding the atmosphere in which clouds and cloud systems form, grow, and dissipate. From the life history and motion of clouds and cloud systems, one obtains direct information on winds and circulation phenomena. Information from Martian cloud observations can be used as direct inputs for the theoretical study of the circulation of the Martian atmosphere; on the other hand, any theoretical models must be able to explain the observed phenomena.

Several observed displacements of moving yellow cloud systems [3] are shown plotted on maps of Mars in Figure 1. The Southern Hemisphere, indicated by the negative latitude numbers, is on the top; the Northern Hemisphere, indicated by the positive latitude numbers, is on the bottom. The time of year for cases A and B corresponds to Southern Hemisphere summer, while the time of year for C, D' and D corresponds to Northern Hemisphere summer on Mars. DeVaucouleurs [3] also mentions two other cloud systems, one in July, 1929 and one in October, 1926 that formed in the same locations and followed similar paths as cloud systems A and B in Figure 1. In addition to these data, deVaucouleurs [4] reported that in the 1958 opposition clouds were again found in the same region and during the same season as shown in cases A and B in Figure 1. In all eight cases except for cases D' and the case of July, 1907, the cloud systems formed at low latitudes of the winter hemisphere. In all cases except D', the clouds moved across the equator, and then started to curve toward the east at middle latitudes of the summer hemisphere. Except for the region of formation, the behavior of these cloud systems is similar to the tropical cyclones in the earth's atmosphere. In the earth's atmosphere, tropical cyclones form at low latitudes of the summer hemisphere and move toward middle latitudes where they generally curve toward the east.

What kind of meteorological conditions would account for the unusual cross-equatorial movement of these cloud systems? Let us consider two possible sequences of events that may account for the observed behavior.

In the first possible sequence, the dust cloud is considered to be the visible manifestation of a cyclonic storm that forms at low latitudes of the winter hemisphere on Mars. Formation and intensification of the storm may be due to the growth of an unstable easterly wave in a fashion similar to what happens in the tropical west Pacific Ocean during winter on earth. Dust is raised by the convergence and upward motion in the center of the storm and is thrown outward by the divergence taking place at the upper levels. The movement of the storm will be controlled by the upper level flow pattern; if the upper level wind field is as illustrated in Figure 2, with broad cross-equatorial flow from winter to summer hemisphere, the storm will be steered toward the equator. Such an upper level flow pattern typically occurs in the earth's atmosphere when a typhoon is formed during the winter season [5]. Figure 3 illustrates the observed upper level flow for such a situation. As

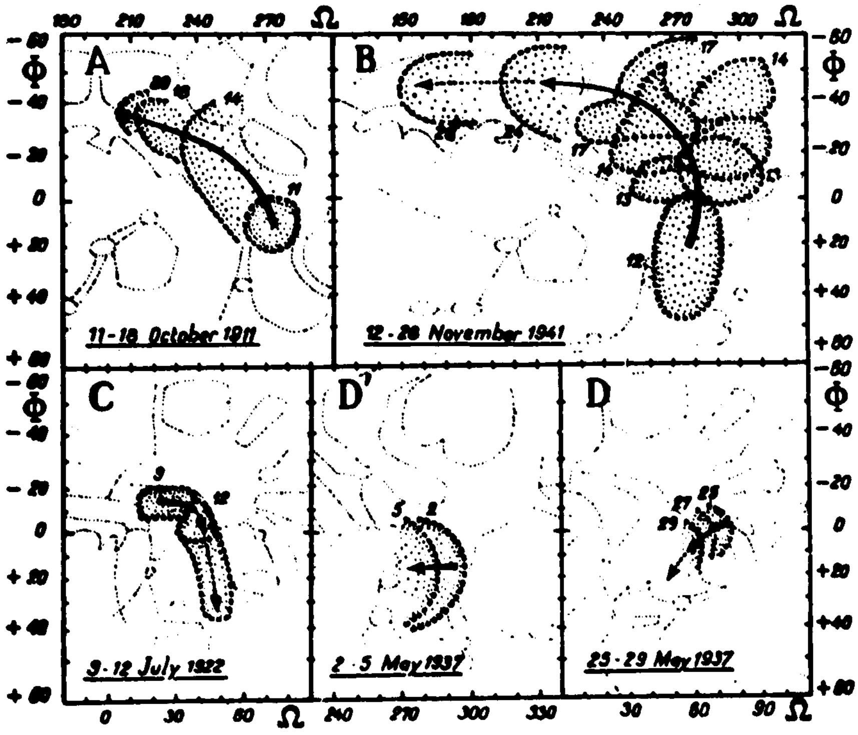

Figure 1. Observed displacement and relative sizes of moving cloud systems (the arrows represent the tracks of the movement of the cloud systems; the heavy dotted curves represent the relative sizes of the cloud systems; the numbers beside the curves represent the dates of the month) [3].

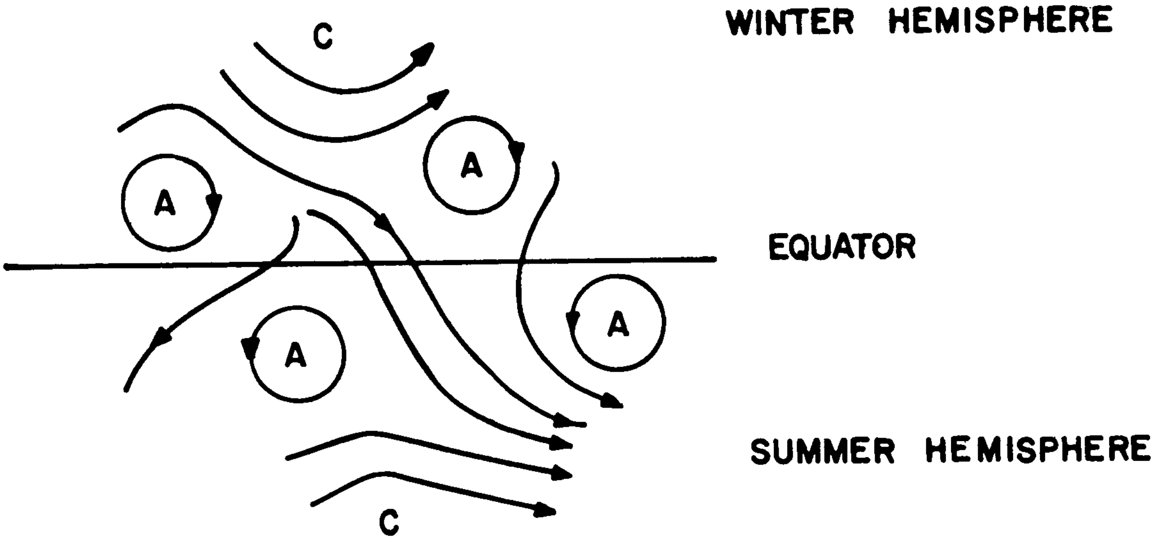

Figure 2. Schematic pattern of relative positions of anticyclonic centers and col areas required for cross-equatorial flow.

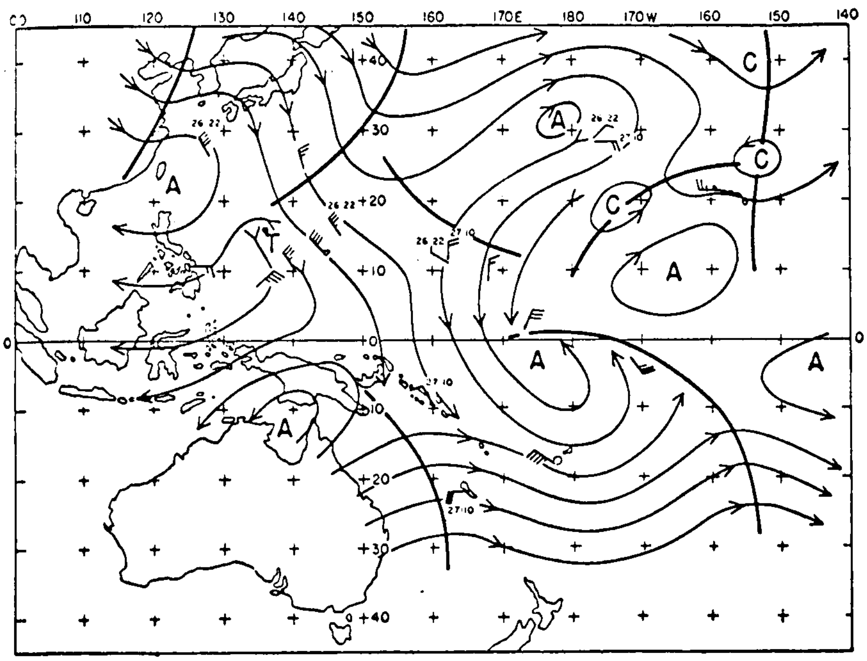

Figure 3. Equatorial 200 mb streamline on earth, 27 September 1945 [6].

the Martian storm approaches the equator, its relative vorticity and, hence, its winds, decrease rapidly. At the equator there may be no storm at all at the surface, but at the upper levels, the divergence field may remain intact as it is advected across the equator. After crossing the equator, the upper level divergence field may induce convergence at the lower levels, and the storm may regenerate itself. As it moves toward higher latitudes, it gains cyclonic vorticity and behaves like a tropical storm on earth.

In the second possible sequence of events, the dust cloud is considered to be generated by the strong winds of a cold, high pressure area moving equatorward. This situation would be similar to wind swept blowing dust conditions found in Northern China. After the Martian cloud system passes the equator, it will follow a trajectory around the periphery of the subtropical high pressure areas on Mars. As in the first sequence of events, the cyclonic vorticity of the dust cloud system will increase as the system moves toward middle latitudes. A true cyclonic storm will form. At the present time, both possible sequences of events must be considered as speculation.

The observed velocities of the cloud systems discussed above are shown in Table 1. The average speed is about 20 to 30 km hr^{-1}. If the yellow cloud systems are manifestations of cyclonic storms, as seems probable for the cases discussed above, their movements would be representative of the storm movement rather than the wind speed within the storm. On earth, for example, the speeds of cyclonic storm movement are less than the speeds of the wind field in which the storm is imbedded. Thus, these observed speeds for Martian yellow cloud systems are probably less than the actual wind velocities within the storms, and also probably less than the wind velocities in the flow field in which the storm is imbedded.

An interesting Martian cloud pattern is shown in Figure 4 [7]. This W-shaped and ribbon-like yellow cloud was observed on 29 August 1956. This pattern is very similar to the jet-stream pattern in the upper terrestrial atmosphere and to wave patterns created in laboratory dishpan experiments that attempt to simulate atmospheric flow patterns. This suggests that a circulation of wave regime exists on Mars and implies that Martian cyclonic centers will move from low latitudes toward and across the jet-stream on a track similar to the ones shown in Figure 1.

Besides these observations of the horizontal motion of the cloud systems, there are interesting observations of the vertical development of clouds. From such observations, additional meteorological information can be deduced. In Figure 5, are shown examples of clouds observed on the morning terminator [3]. It was reported that the dull grey cloud, sloping upward from the surface, extended more than 40 degrees in longitude and remained visible for three hours with a top at about 27 km. In the second case, although the high cloud separated from the surface, it still remained tilted like a frontal cloud, and extended to a height of at least 30 km. These observations show that these clouds extend to very high altitudes and have sloping surfaces just like a schematic terrestrial front.

TABLE 1

OBSERVED VELOCITIES OF SOME MOVING CLOUD SYSTEMS*
(based on [3] and [4])

Case	Date	Displacement (km)	Velocity km hr^{-1}	Remarks
A	11-14 October 1911	2000 - 2500	30	Yellowish-
	14-18 October 1911	1000 - 1500	13	white color
	18-20 October 1911	0 - 500	5	
B	12-13 November 1941	1200 - 1800	60	Yellowish
	13-17 November 1941	2000 - 2500	23	
	17-24 November 1941	1500 - 2000	10	
	24-28 November 1941	1500 - 2000	18	
C	9-12 July 1922	1200 - 1500	18(36)	Creamy-white
D'	2-5 May 1937	720	10	Bright-white
D	25-29 May 1937	1200	13	Bright-white
E	25-26 October 1926	1800 - 2400	90	Yellowish
F	14-15 October 1958	1000 - 1200	50	Bright-white
	15-16 October 1958	1500 - 1800	70	
	16-17 October 1958	500 - 800	24	

*Cases A, B, C, D', and D corresponding to that of Figure 1.

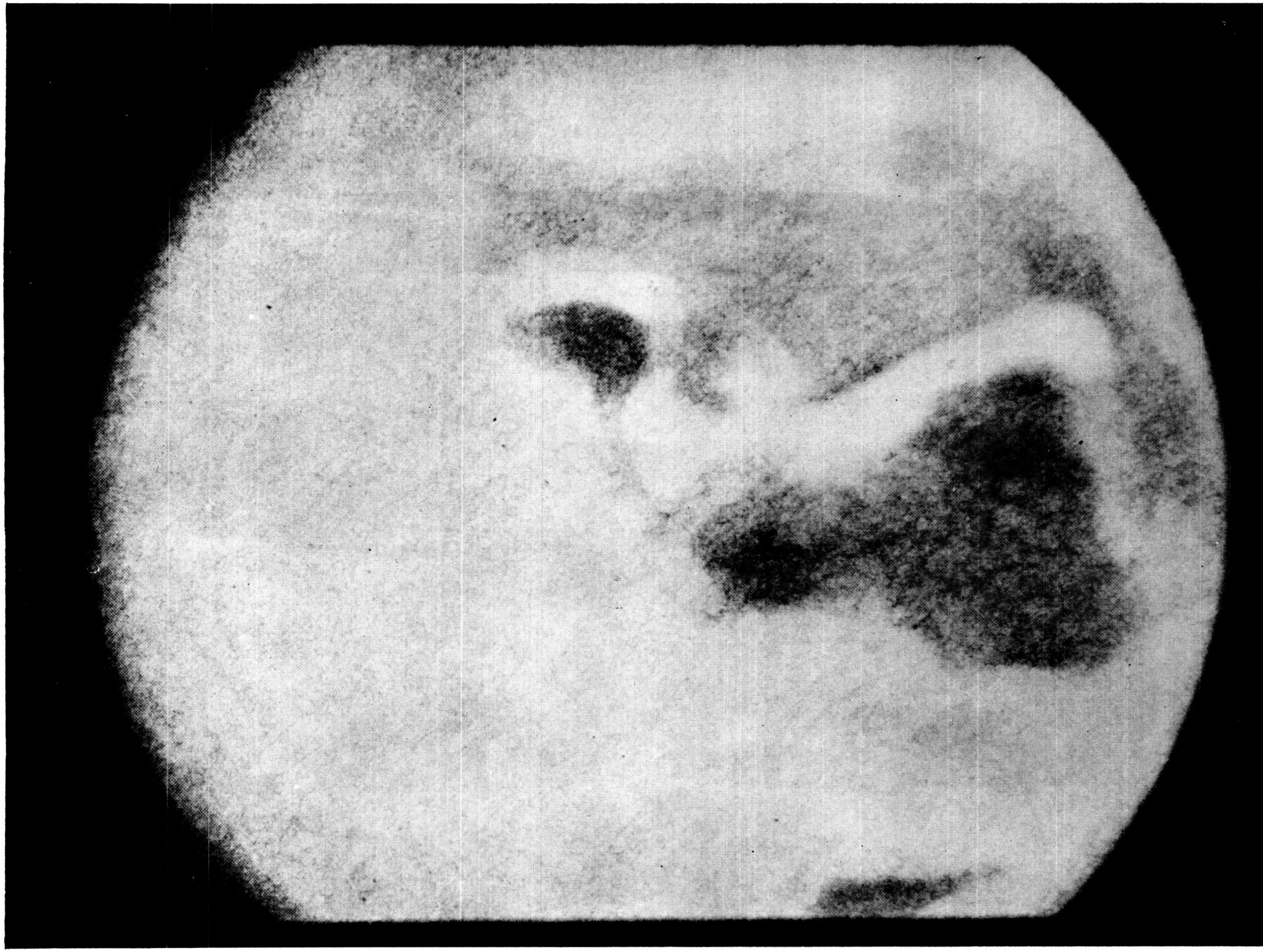

Figure 4. W-shaped cloud on Mars [7].

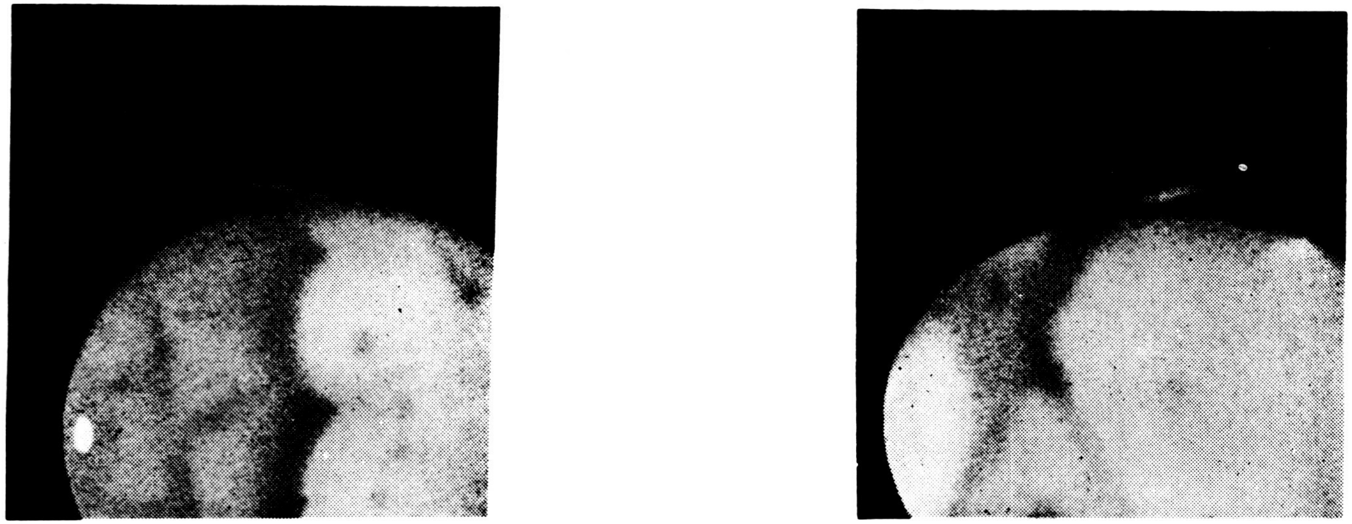

Figure 5. Observations of clouds on the terminator [3]. Case 1. The dull-grey cloud extended over more than 40 degrees in longitude and remained visible for 3 hours with top at about 27 km. Case 2. The dark yellow cloud, which appeared separated from the surface, remained tilted like a frontal cloud, and extended to a height of at least 30 km [3].

If the clouds are frontal clouds, then the heights to which they extend are indicative of the height of the tropopause on Mars. The observations suggest a tropopause height of about 30 km at low latitudes. If this interpretation is correct, the height of the tropopause on Mars is at least twice that of earth.

In summary, we make the following inferences on Martian circulation features, based upon our interpretation of the observations of Martian clouds.

(1) There is a tendency for dust cloud systems to form at low latitudes of the winter hemisphere. These dust clouds are probably formed as a result of cyclonic storms or the strong winds of a polar outbreak. These cloud systems move across the equator, enter the opposite hemisphere, and curve toward the east at mid-latitudes. Curvature toward the east suggests the presence of Martian subtropical high pressure areas similar to those on the earth. The transequatorial drift of these cloud systems is indicative of meridional flow in the upper part of the Martian equatorial atmosphere. Similar flow conditions occur in the Philippine Islands area on the earth during late fall or winter.

(2) If the observed movements of these yellow clouds are representative of storm movements on Mars, then the average drift velocity — about 20-30 km hr^{-1} — of the cloud systems is probably less than the speed of the broad current in which the storm is imbedded.

(3) It appears that the general circulation on the planet Mars is in the wave regime, as is the earth's.

(4) Cloud systems analogous to frontal cloud systems on earth exist on Mars. Thus, fronts may be a typical meteorological phenomena of the Martian atmosphere. These cloud systems can extend to 30 km; if this height is interpreted as the height of the Martian tropopause then the Martian tropopause is twice as high up as the earth's tropopause.

These observational inferences will be compared with theoretical inferences which are made in the next few sections.

2. THE SLOPE OF A FRONT

The cloud pictures shown in Case 1 of Figure 5 resemble the terrestrial front, which is characterized by its coverage of a large area and by the sloping, relatively narrow cloud band above its surface. Furthermore, the fact that these phenomena are observed on the terminator suggests that they are not convective clouds for the following reasons. Convection must be weakest at the time of sunrise on a desert-like planet. Also, the sloping narrow cloud band bears no resemblances to a convective cloud with strong vertical development as observed on the earth.

The slope of a terrestrial front at middle latitudes is about 1/50 to 1/100. The slopes of the Martian cloud systems in Figure 5 are much greater than 1/50 to 1/100. Another interesting feature of the cloud systems in Figure 5 is that they are located in the equatorial region. On earth, fronts and frontal cloud systems are usually not found in equatorial regions.

Now let us determine if the observed slope of the cloud system is compatible with the slope of a front as determined from frontal theory. The conventional formula with geostrophic approximation, for a frontal slope is only applicable to cases at middle latitudes. For low latitudes, the geostrophic approximation is not valid, and one must include the advective terms in the equation for the slope of a front.

The expression for the slope of an east-west front at low latitudes, can be written as

$$\tan \alpha = - \frac{\rho_1(U\frac{\partial U}{\partial x} - fV)_1 - \rho_2(U\frac{\partial U}{\partial x} - fV)_2}{+(\rho_1 V_1 - \rho_2 V_2)\frac{\partial f}{\partial \varphi} + g(\rho_1 - \rho_2)}$$

where α = the angle between the frontal surface and the horizontal surface.
ρ = density.
f = Coriolis parameter.
φ = latitude.
g = gravitational constant.
x,y = the coordinate system in which x and y are in the direction normal and parallel to the surface front in the horizontal surface respectively. (Assume x points toward north and y points west).
U,V = velocities along x- and y-direction respectively,

and,

subscripts 1 and 2 denote the cold and warm air masses respectively.

At low latitudes, one would expect a very small horizontal density contrast because of small temperature gradients, and, therefore, one can assume that

$$\tan \alpha \cong - \frac{f(V_1 - V_2) - (U\frac{\partial U}{\partial x})_1 - (U\frac{\partial U}{\partial x})_2}{\frac{\partial f}{\partial \varphi}(V_2 - V_1)} .$$

Furthermore, we may assume that the mean warm air velocity normal to the front, U_2, is negligibly small and that the mean velocity gradient in the cold air is constant, say c. With these approximations, the slope of the front can be written as

$$\tan \alpha \cong \tan \varphi + \frac{c|U_1|}{2\omega \cos \varphi(V_1 - V_2)} .$$

The value c may be estimated as

$$c \cong \frac{\Delta U_1}{\Delta x} = \frac{10 \text{ m sec}^{-1}}{2000 \text{ km}} = \frac{1}{2} \times 10^{-5} \text{ sec}^{-1} .$$

Further assume

$$|U_1| = 40 \text{ m sec}^{-1}.$$

$$V_1 - V_2 = 10 \text{ m sec}^{-1}.$$

$$2\omega = 1.5 \times 10^{-4} \text{ sec}^{-1}.$$

$$\varphi = 5^{\circ} \text{ latitude.}$$

Then

$$\tan \alpha \approx 0.22$$

and

$$\alpha \approx + 12^{\circ}.$$

This value agrees well with the slope of clouds shown in Figure 5. The assumptions made in the derivation of this value are all reasonable ones, and, therefore, it must be concluded that the observed cloud systems may be representative of Martian frontal phenomena.

The reason for the high frontal slope, compared to mid-latitude frontal slopes in the earth's atmosphere, is that the Martian front is located at low latitudes, where density contrasts are presumably small and the effect of advection is more important than the Coriolis effect.

If these clouds are indeed frontal clouds, they are formed by condensation processes during up-glide motion along the front. Hence, they would be composed of a condensable substance such as water vapor rather than dust. The cloud of Case 1 of Figure 5 had a dull-grey color, in agreement with an ice crystal composition, the cloud of Case 2 was dark yellow in color, which suggests the possibility of dust mixed with ice particles.

The fact that cloud system 1 of Figure 5 disappeared three hours after sunrise leads further credence to an ice crystal composition; with the rise of the sun and solar heating, a thin ice crystal cloud would evaporate, whereas a dust cloud would not be affected.

Although rare, frontal phenomena extending to equatorial regions on earth also occur. They usually occur during the winter as the result of an unusually cold outbreak moving far south. These occurrences are limited to the huge land masses of the Northern Hemisphere, especially Asia. For example, it is not too unusual for a frontal cloud cover to extend from Indo-China to the Yellow River region during winter — that is, from 10°N to 35°N, which is comparable to the cloud system of Case 1 in Figure 5. Since such frontal phenomena occur over huge land masses on earth, and since the Martian surface is one huge land mass, there is further justification for suspecting that the observed cloud systems of Figure 5 are indeed frontal clouds.

3. THE CIRCULATION REGIME

For planetary scale motion, the terrestrial atmosphere is in a state of quasi-geostrophic balance. It is of interest to determine whether the large scale Martian atmospheric motions are in a similar state of quasi-geostrophic balance. The criterion for such a balance is based on a non-dimensional number — the Rossby number, R_o — which is the ratio of the inertial force to the Coriolis force for a given flow of a rotating fluid. It may be written as

$$R_o = \frac{U}{fL}$$

where U is a characteristic velocity, f the Coriolis parameter and L a characteristic length. This number was first used explicitly in meteorology for the scale analysis of atmospheric motion by Charney [8]. In the terrestrial, R_o is about 1/10 or less, which means that the inertial acceleration — U^2/L — is one order of magnitude smaller than the Coriolis acceleration — fU. As a consequence, the equation of motion may be simplified by neglecting the inertia term, i.e., the Coriolis force balances just the pressure gradient force. The smaller the Rossby number, the better the correspondence of the motion to geostrophic conditions. The Rossby number may be expressed in different ways. If the inertia force due to the thermal wind is used, then it is called "thermal Rossby number." This number was originally developed as a basic modeling criterion in laboratory experiments of the general circulation of the atmosphere. The laboratory fluid annulus experiment is dynamically similar to the atmosphere of a rotating planet because both circulations are thermally driven. Since both the atmosphere and laboratory experiment are thermally driven, the thermal Rossby number can be applied in either case. When the thermal Rossby number is of the order of 0.1 or less, the hydrodynamic equation of a rotating fluid can be simplified to the geostrophic wind equation [9]. Furthermore, we can use the thermal Rossby number to investigate the circulation regime of the rotating fluid — that is, to determine whether the general circulation is in a symmetrical or wave regime. By estimating the thermal Rossby number of the Martian atmosphere, we can determine whether the motions are geostrophic and the type of circulation regime.

The laboratory model version of the thermal Rossby number, R^*_{oT}, may be written as [10]

$$R^*_{oT} = \frac{g \cdot \alpha \cdot (\Delta_r T) \cdot H}{(r_o \Omega)\, f\, \Delta r}$$

where g = gravitational acceleration.
α = coefficient of volume expansion of the fluid.
$\Delta_r T$ = radial temperature difference.
H = depth of the fluid.
r_o = maximum radius.
Ω = absolute rotation of the coordinate system.
f = Coriolis parameter.
Δr = width of the fluid layer in the radial direction.

To apply this expression to an atmosphere, we must substitute for $\Delta_r T$ the temperature difference between equator and pole; for r_o the radius of the planet; for Δr the distance between equator and pole; and for H the scale height of the atmosphere. For the Mars' atmosphere the following numerical values are appropriate:

$$g = 3.8 \text{ m sec}^{-2}$$

$$\alpha = 1/200$$

$$H = 20 \text{ km} = 2 \times 10^{4} \text{ m}$$

$$r_o = \text{radius} = 3.4 \times 10^{6} \text{ m}$$

$$\Omega = \frac{2\pi}{24.5 \text{ hour}} = 6.4 \times 10^{-5} \text{ sec}^{-1}$$

$$f = 0.9 \times 10^{-4} \text{ sec}^{-1}$$

$$\Delta r = \frac{\pi}{2} r_o = 5.1 \times 10^{6} \text{ m.}$$

For a mean surface temperature difference between the equator and pole, 30C, as given by Ohring, Tang, and DeSanto [11] and with the assumption that the vertically averaged temperature difference between the equator and pole is two thirds of the above quantity, i.e. 20C, we obtain R^*_{oT} = 0.073. For a temperature difference of 22C, computed from Mintz's [2] temperature distribution, the thermal Rossby number is increased by 10%. These values are compatible with the values of the thermal Rossby number for the terrestrial atmosphere, which are shown in the following table (After Fultz et al., [10], based upon Mintz's [12] data for 1949):

Season	R^*_{oT}
Summer	0.027
Winter	0.054

Since R^*_{OT} in the Martian atmosphere is of the same order of magnitude as in the earth's atmosphere, the mean annual atmospheric motion on Mars is probably in quasi-geostrophic equilibrium, as on the earth. With the computed thermal Rossby number, R^*_{OT}, and the rotation parameter, $G^* = F^{-1} = g(r_o\Omega^2)^{-1}$, ($F$ = Froude number), one can determine whether the circulation regime is a symmetrical or wave type; and, if it is a wave regime, one can determine the preferred wave number. Transition curves for different wave numbers based on the experimental results of Fultz et al. [13] are shown in Figure 6. The coordinates are $(G^*)^{-1}$ and R^*_{OT}. Based on the physical parameters of the planet Mars, $(G^*)^{-1}$ is approximately 4.3×10^{-3}. From $(G^*)^{-1}$ and the previously computed R^*_{OT}, we obtain the result that the Martian atmosphere is in a wave regime with a preferred wave number of about four, which is close to the wave number for the terrestrial general circulation.

On the other hand, Mintz [2] reached a different conclusion. From consideration of thermal equilibrium he derived a criterion for the stability of a symmetrical circulation regime as follows:

$$\Delta Q \lessgtr \Delta Q_{crit.} \qquad \begin{matrix} \text{stable} \\ \text{unstable} \end{matrix}$$

where ΔQ = actual total poleward heat transport,

$\Delta Q_{crit.}$ = critical value of total poleward heat transport.

If the symmetrical circulation regime is unstable, a wave type regime will prevail.

The critical value of the total poleward heat transport can be written as

$$\Delta Q_{crit.} = \frac{\pi\, g\, \kappa\, R^*\, \tilde{T}_2\, \mu_2\, S^2}{\sqrt{2}\, m\, \Omega^2}$$

where g = gravitational acceleration.

$\kappa = \dfrac{R^*}{m\, c_p}$ = Poisson constant.

R^* = universal gas constant.

$\tilde{T}_2$ = mean temperature at the middle level of the atmosphere.

μ_2 = coefficient of eddy viscosity at the middle level of the atmosphere.

$S = (1 - \dfrac{\gamma}{\gamma_d})$ = static stability.

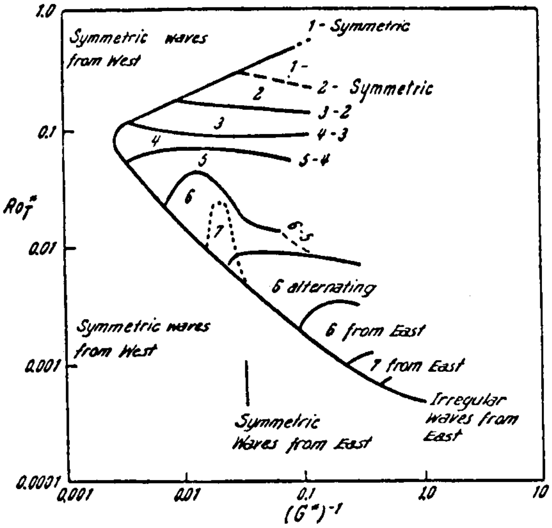

Figure 6. Experimental transition curves for fluid in a tall annulus [13].

$\gamma_d = - \frac{g}{c_p}$ = adiabatic vertical temperature gradient.

m = mean molecular weight.

c_p = specific heat at constant pressure.

Ω = angular velocity of the planet.

Using

$\Omega = 7.1 \times 10^{-5}$ sec $^{-1}$.

$g = 3.8$ m sec^{-2}.

$\mu_2 = \mu_2$(earth) $= 220$ g sec^{-1} cm^{-1}.

$\frac{R^*}{m} = 297$ m^2 sec^{-2} deg^{-1}.

$\gamma_d = -3.8$C km^{-1}.

Mintz obtains

$$\Delta Q_{crit.} = 0.17 \times 10^{12} \text{ kj sec}^{-1}.$$

ΔQ, based on a computation of the mean annual radiation budget, is

$$\Delta Q = 0.12 \times 10^{12} \text{ kj sec}^{-1}.$$

Therefore, he concludes that, in the mean for the year on Mars, $\Delta Q < \Delta Q_{crit.}$, and the circulation is in the symmetrical regime and will be dynamically stable.

It is noticed that $\Delta Q_{crit.}$ is directly proportional to the coefficient of eddy viscosity, μ_2, at the middle level of the atmosphere. The value of μ_2 assumed by Mintz is 220 g sec^{-1} cm^{-1} and is adopted from Palmén [14]. However, later Palmén [15] assumed $\mu = 100$ g sec^{-1} cm^{-1} in the computation for the mean frictional dissipation in the layer 1-12 km in the terrestrial atmosphere. Other workers, such as Schmidt [16], Rossby and Montgomery [17], Haurwitz [18] and recently Palmén [15] have assumed that μ is approximately 100 g sec^{-1} cm^{-1} in the free atmosphere or even in the planetary boundary layer on the earth as shown in Table 2.

TABLE 2

DYNAMIC COEFFICIENT OF EDDY VISCOSITY IN VERTICAL DIRECTION ASSUMED BY VARIOUS WORKERS

Source	Dynamic Coefficient of Eddy Viscosity in Vertical Direction μ(g sec^{-1} cm^{-1})
Schmidt [16]	100
Rossby & Montgomery [17]	50
Lettau [19]	70 (gradient wind level)
Haurwitz [18]	116
Palmén [14]	220
Palmén [15]	100

If $\mu = 100$ g sec^{-1} cm^{-1} is used instead of 220 g sec^{-1} cm^{-1} of Mintz's criterion, then $\Delta Q > \Delta Q_{crit.}$ and the conclusion is reached that the general circulation on Mars is in a wave regime in the mean for the year. This is in agreement with the conclusion based upon the thermal Rossby number and the rotation parameter approach.

4. MEAN ZONAL AND MERIDIONAL WIND VELOCITIES OF GENERAL CIRCULATION AND MAXIMUM SURFACE WIND VELOCITY

Circulation Estimates Based upon Haurwitz's Model

The model used below in the computation of the mean zonal and meridional wind velocity on Mars is based on a general theoretical model of thermally driven large scale motion on a rotating planet. In order to use this model, one has to know the difference in the rate of temperature change between the equator and pole. This can be obtained from the annual radiation budget. The most recent available annual radiation budget for Mars, as computed by Ohring et al. [20], gives the total difference of the rate of heating per unit area between the equator and pole:

$$\Delta[S(1 - A) - W] = 0.04 \text{ cal min}^{-1} \text{ cm}^{-2}$$

where
- Δ = difference of the quantities in the brackets between the equator and pole.
- S = incident solar radiation.
- W = outgoing long-wave radiation.
- A = planetary albedo.

By the first law of thermodynamics

$$\frac{\delta Q}{dt} = \frac{\Delta[S(1 - A) - W]}{\Delta p/g} ,$$

where $\delta Q/dt$ is the difference in the heating rate per unit mass between equator and pole, and Δp is surface pressure. The difference in the rate of temperature change between equator and pole, α, will be

$$\alpha = \frac{g \cdot \Delta[S(1 - A) - W]}{c_p \cdot \Delta p} .$$

With the latest finding of surface pressure of Mars, Δp = 25 mb. Let c_p = 0.239 cal g^{-1} deg^{-1} and g = 380 cm sec^{-2}. Then

$$\alpha = 4.2 \times 10^{-5} \text{K sec}^{-1}.$$

If the radiation budget of Mintz's [2] model is used, then $\Delta[S(1-A) - W] = 0.023$ cal min^{-1} cm^{-2}, which is about 60% of Ohring and al.'s value. The difference is due to different assumptions in the estimation of the Martian radiation budget. The corresponding α will be also reduced to

$$\alpha = 2.43 \times 10^{-5} K\ sec^{-1}.$$

Both these values will be used to compute the mean wind velocities for Mars.

The differential solar heating rate, as expressed by the difference in the rate of temperature change between equator and pole, will drive a direct circulation between equator and pole. The effect of rotation of the planet is deflection of the motion so that the meridional velocity component is reduced, while a larger zonal velocity component develops. For a continuous input of solar energy into the system, dissipative processes such as friction have to be considered in the model so that the mean motion can approach a steady state asymptotically. Taking all these important factors into consideration, Haurwitz [21] developed a simple model for a thermally driven circulation for a general, rotating fluid system.

The steady state solutions for mean meridional and zonal wind, as approached asymptotically, are

$$V = \frac{1}{2}\left[\frac{\alpha R\ \ell n(\frac{p_o}{p_1})}{k(1 + \frac{f^2}{k^2})}\right]^{1/2}$$

$$U = \frac{f}{k} V$$

respectively, where

p_o, p_1 = pressure at the surface and top of the circulation system respectively.
α = difference of the rate of change of temperature between the equator and pole.
R = gas constant of the atmosphere of planet.
k = coefficient of friction.
f = Coriolis parameter.

The most important factor in the solution of the mean wind velocities is the quantity, α, i.e., the difference of rate change of temperature between equator and pole, caused by differential heating, which may be considered as the energy source of the motion of the atmosphere. In this model of the general circulation, the coefficient of friction is also very important and the result is quite sensitive to the value of the coefficient of friction. The numberical results are obtained and compared with the observations on earth in the following paragraphs.

Now, let us assume a surface pressure 25 mb, in agreement with recent observations, and assume the circulation can reach the level corresponding to the level at the top of seventy-five per cent of the atmosphere, i.e., 6.25 mb (See Section 5), then

$$\ell n \frac{p_o}{p_1} = \ell n \; 4 = 1.386 \; .$$

The coefficient of friction in the Martian atmosphere is probably less than in the earth's atmosphere because of the lower Martian atmospheric densities. We shall assume a value of k equal to one half of the value used by Haurwitz [21] for the earth's atmosphere. With $k = 5 \times 10^{-6}$ sec^{-1} and $\alpha = 4.2 \times 10^{-5}$K sec^{-1},

$$V \cong 1.5 \text{ m sec}^{-1} \qquad \text{and} \qquad U = 30.0 \text{ m sec}^{-1} \; .$$

Computations have also been performed for surface pressures of 85 mb and with Mintz's [2] radiation budget. In Table 3, the theoretical calculations are compared with each other, with observed Martian velocities, and with theoretical and observed values for the earth.

It is seen that the computed values of velocity for earth are reasonably well in agreement with observation. The computed values for Mars in the 25 mb case indicate that the average zonal wind velocities on Mars are about three to four times as high as the computed zonal wind velocity on earth; and that the average meridional wind velocities on Mars are about the same as the computed average meridional wind velocity on earth. In the 85 mb case, the computed zonal wind velocities are almost twice those of the earth.

It is also seen in Table 3 that the computed average wind velocities for Mars are larger than the observed average speed of moving yellow cloud systems. If the yellow clouds are manifestations of atmospheric vortices, they will move at a speed lower than the average speed of the current in which they are imbedded. Therefore, the observed average velocity of the yellow clouds is not representative of the average velocity of the atmosphere.

That the computed average wind speeds on Mars are higher than those computed for the earth may be attributed to:

(1) A larger difference in radiational temperature change between equator and pole on Mars than on earth.

TABLE 3

THE COMPUTED AND OBSERVED MEAN WIND VELOCITIES FOR EARTH AND MARS FOR VARIOUS CONDITIONS

Planets	Important Parameters and Items on Which the Computations of Velocities are Based	Velocity ($m\ sec^{-1}$) Zonal U	Meridional V
Earth	Computed value [21]	8.15	0.82
	Observed yearly mean value for earth's southern hemisphere [22]	8.05	0.37*
Mars	Computed value		
	1. Surface pressure = 25 mb		
	a. Radiation budget $\Delta[S(1-A)-W] = 0.04$ cal $cm^{-2} min^{-1}$	30.0	1.5
	b. Radiation budget $\Delta[S(1-A)-W] = 0.023$ cal $cm^{-2} min^{-1}$	23.0	1.1
	2. Surface pressure = 85 mb		
	a. Radiation budget $\Delta[S(1-A)-W] = 0.04$ cal $cm^{-2} min^{-1}$	16.2	0.8
	b. Radiation budget $\Delta[S(1-A)-W] = 0.023$ cal $cm^{-2} min^{-1}$	12.6	0.6
	Average velocity of moving yellow cloud systems [23]	9	

*60-70°S latitude.

(2) A smaller dissipation of kinetic energy by friction on Mars than on earth.

Both of these circumstances are indirectly due to the fact that the Martian surface pressure is lower than the earth's.

The computed Martian winds are higher in the 25 mb case than in the 85 mb case. With lower pressures — hence, lower masses — the radiational heating rates per unit mass are increased, resulting in a greater latitudinal difference of rates of atmospheric temperature change. This leads to higher velocities.

The Estimation of Maximum Surface Wind on Mars

The maximum surface wind speed, such as in a storm, must be larger than these mean values. The strongest surface wind velocities in the terrestrial atmosphere are found in tornadoes and typhoons or hurricanes. The maximum surface wind velocity on Mars would also be found in the corresponding vortex, if there are such dynamic phenomena on the planet. (Undoubtedly, on Mars, these would be dry hurricanes, tornadoes, etc.). A formula for the maximum surface wind velocity for a convective vortex, such as a tropical storm, was derived by Kuo [24].

$$V_{max.} = \gamma \, RT_o \left\{ \frac{2}{\gamma - 1} \left[1 - \left(\frac{P_c}{P_o}\right)^{\kappa} \right] \right\}^{1/2}$$

where $V_{max.}$ = maximum tangential velocity.
γ = the ratio of specific heat at constant pressure to that at constant volume = C_p/C_v.
R = gas constant for the atmosphere.
P_c = surface pressure at center of the vortex (central pressure)
P_o = surface pressure at a distance where the wind velocities are nil (surrounding pressure).
κ = Poisson constant= $(C_p - C_v)/C_p$.
T_o = surface temperature.

In the case of a terrestrial typhoon or hurricane, the lowest surface central pressure, P_c, is about 950 mb and the surrounding surface pressure P_o is 1000 mb. With a sea surface temperature, T_o, of 290K the computed maximum velocity is 93.7 m sec^{-1}, which is comparable to the observed wind speed in a hurricane. On Mars, the average surface temperature in the equatorial area is about 280K. The central pressure of a vortex corresponding to a tropical storm has not been determined. However, a disturbance of surface pressure in a range of 2 to 3 mb in the equatorial area of Mars may be possible. If this central pressure is 2 mb lower than the surrounding surface pressure (P_o= 25 mb), the corresponding maximum surface wind velocity in the vortex is approximately

$$V_{max.} = 114 \text{ m sec}^{-1}.$$

For $P_c = 3$ mb, $V_{max.}$ is approximately

$$V_{max.} = 140 \text{ m sec}^{-1}.$$

Both wind speeds are larger than the maximum wind speed in hurricanes on earth. In the governing equation for the maximum velocity, the dominating factor is the ratio of the central pressure to the surrounding pressure. Since the latest estimate of surface pressure on Mars is about 25 mb, a slight disturbance will create a severe storm. A central pressure of 23.75 mb on Mars leads to the same maximum wind as a central pressure of 950 mb on earth. Whether pressure drops of 1.25 to 3 actually occur on Mars is unknown.

The observed large diurnal variation of surface temperature on Mars are indicative of strong heating of the surface during the daytime. Under such conditions of strong surface heating, strong convective phenomena may take place. Observations of yellow clouds extending to heights of 10 km and higher are suggestive of deep convective storms. Pressure drops of 1 to 3 mb in such storms are not inconceivable.

Let us attack the problem of estimating the maximum surface wind on Mars from a slightly different approach. The gradient wind formula relates the wind speed to the pressure gradient in a cyclone. Some of the yellow cloud systems discussed in Section 1 are suggestive of tropical cyclones. The yellow cloud system observed on 11 October 1911 had a radius of 500 km. If we assume a value for the pressure drop (or gradient) in this cloud system, we can use the gradient wind formula to compute the average wind velocity. We assume a pressure drop of 3 mb, a surface atmospheric density of 5×10^{-5} g cm^{-3}, and a latitude of 15° for the location of the storm. The mean wind velocity from the gradient wind formula is

$$\bar{V} = \frac{\bar{r}f}{2}\left[\left(1 + \frac{4}{\bar{r}f^2}\frac{1}{\rho}\frac{\partial P}{\partial r}\right)^{\frac{1}{2}} - 1\right]$$

where
$\bar{V}$ = mean wind velocity.
r = distance from center.
$\bar{r}$ = mean radius = $r_o/2$.
r_o = radius of the system.
f = Coriolis parameter.
ρ = density.
P = pressure.

With the parameters assumed above

$$\overline{V} = 52 \text{ m sec}^{-1}.$$

Since it may be assumed that the wind increases roughly at a linear rate from zero at the outside of the storm to a maximum near the center, we may estimate the maximum wind to be double the computed average wind, or 104 m sec^{-1}. This is a reasonable agreement with the maximum wind of 140 m sec^{-1} for a 3-mb pressure drop estimated with the maximum wind speed formula for a convective vortex.

5. THEORETICAL ESTIMATES OF THE PROFILE OF LARGE SCALE VERTICAL VELOCITY AND THE VERTICAL EXTENT OF CLOUDS ON MARS

In this section, we would like to obtain some estimates of the magnitudes and variation with altitude of the vertical velocities associated with Martian atmospheric waves. Knowledge of the height variation of the vertical velocity will enable us to estimate the maximum vertical velocity and the maximum heights that frontal type clouds can reach.

Since the large scale motion on Mars is generally quasi-geostrophic in nature as discussed in Section 2, and, further, the atmosphere may be assumed quasi-static, the vorticity and thermal equation for an adiabatic and frictionless atmospheric system (See, for example, [25]) can be applied to the Martian atmosphere and written as

$$\frac{\partial}{\partial t} \nabla^2 \Phi + \frac{1}{f} J(\Phi, \nabla^2 \Phi) + \beta \frac{\partial \Phi}{\partial x} = f^2 \frac{\partial \omega}{\partial p} \tag{5.1}$$

$$\frac{\partial}{\partial t} \frac{\partial \Phi}{\partial p} + \frac{1}{f} J\left(\Phi, \frac{\partial \Phi}{\partial p}\right) = \frac{S\omega}{p} \tag{5.2}$$

where $\Phi = gz$.

$S = RT \dfrac{\partial \ln \theta}{\partial p}$.

J = Jacobian.

p = pressure.

∇^2 = two-dimensional Laplacian.

f = Coriolis parameter.

$\beta = \dfrac{\partial f}{\partial y}$.

$\omega = \frac{dp}{dt}$.

t = time.

y = distance, with positive direction toward north.

Since we are interested in estimating the profile of the vertical velocity in a large scale disturbance, a harmonic wave solution for Φ, the geopotential, may be assumed as follows:

$$\Phi = C(p + b)\gamma + B(p) \exp [i(k_1x + k_2y - \sigma t)] \tag{5.3}$$

where $C = -f \frac{\partial u}{\partial y}$.

b = a constant.

k_1, k_2 = wave numbers in the x- and y-direction respectively.

σ = wave frequency.

Following Haltiner and Martin [25], we eliminate $(\partial^2\Phi)/(\partial t \partial d)$ from equation (5.1) and (5.2), substitute (5.3) for Φ, and use the geostrophic approximation to obtain

$$\frac{f^2}{k^2}\frac{\partial^2\omega}{\partial p^2} + \frac{S}{p}\,\omega = - 2 \frac{\partial u}{\partial p}\, fv + \frac{\beta f}{k^2}\frac{\partial v}{\partial p} \tag{5.4}$$

where $k^2 = k_1^2 + k_2^2$. This is a simplified "Omega equation" which will be used here to determine the vertical profile of ω. Now let

$$s = 2(\frac{p}{P_o})^{\frac{1}{2}}$$

where P_o is the pressure assumed at surface level. Further assume the velocity profile to be similar to that on earth.

$$u = u_o(1 - \frac{p}{P_o}) ,$$

$$v = v_o(0.4 - \frac{p}{P_o})$$

where u_o, v_o = amplitude of the zonal and meridional velocities. This means that the zonal velocity increases with height from zero at surface to u_o at $p = 0$ and the meridional velocity increases with height from $-0.6v_o$ at the surface to zero at 0.4 P_o and to 0.4 v_o at the top of the atmosphere. If ω is assumed in phase with v in zonal direction, then Equation (5.4) can be written as

$$\frac{d^2\omega}{ds^2} - \frac{1}{s}\frac{d\omega}{ds} - \gamma\omega = \frac{Ds^2}{4}\left(0.4 - \frac{s^2}{4}\right) - \frac{G}{4}s^2 \qquad (5.5)$$

where

$$\gamma = \frac{R^2|S|P_o}{f^2}$$

$$G = \frac{\beta\, P_o\, v_o}{f}$$

$$D = \frac{2k^2\, P_o\, u_o\, v_o}{f} \quad .$$

The right-hand side of Equation (5.5) may be written as

$$Es^2 - Fs^4$$

where

$$E = \frac{0.2\, k^2\, P_o\, u_o\, v_o}{f} - \frac{\beta\, P_o v_o}{4f}$$

$$F = \frac{k^2\, P_o\, u_o\, v_o}{8f} \quad .$$

If the boundary conditions for the vertical velocity, ω, are assumed $\omega = 0$ at $P = 0$ and $P = P_o$, then the complete solution of Equation (5.5) is

$$\omega = \frac{-2}{\gamma}\,\frac{[4F + (-E + \frac{8F}{\gamma})]}{I_1\,(2\gamma^{\frac{1}{2}})}\; sI_1(\gamma^{\frac{1}{2}}s) + \left(\frac{-E}{\gamma} + \frac{8F}{\gamma^2}\right)s^2 + \frac{F}{\gamma}\,s^4 \qquad (5.6)$$

However, in general, for medium or short wave disturbances, the last term in Equation (5.5) can be neglected for, in these cases, G is small. Then, the solution becomes

$$\omega = - \frac{D}{\gamma I_1(2\gamma^{\frac{1}{2}})} (0.3 + \frac{1}{\gamma}) s I_1 (\gamma^{\frac{1}{2}} s) + \frac{D}{16\gamma} s^4 + \frac{D}{2\gamma^2} (1 - 0.2\gamma) s^2 \qquad (5.7)$$

For horizontal wind velocities at the top of a wave disturbance we assume

$$u_o = v_o = 40 \text{ m sec}^{-1} .$$

Furthermore, we let

$$f = 10^{-4} \text{ sec}^{-1} .$$

$$R = 3 \times 10^2 \text{ m}^2 \text{ sec}^{-2} \text{ day}^{-1}.$$

$$k = \frac{2\pi}{4000} \text{ km}^{-1} .$$

$$\frac{T}{\theta} \frac{\partial \theta}{\partial p} = \frac{-4.0}{100} \text{ C mb}^{-1}.$$

$$|S| = 12 \text{ m}^2 \text{ mb}^{-1} \text{ sec}^{-2} \text{ deg}^{-1} .$$

$$\rho = 10^{-3} \text{ g cm}^{-3} \text{ at surface.}$$

$$\rho = 0.6 \times 10^{-3} \text{ g cm}^{-3} \text{ at 600 mb.}$$

The computed profile of ω in front of a trough in the earth's atmosphere, using the above values, is shown in Figure 7. ω reaches a negative maximum near 650 mb, decreases to zero at about 250 mb, and reaches a positive maximum at about 80 mb. A negative value of ω corresponds to upward vertical motion, a positive to downward vertical motion. The computed maximum vertical velocity

$$w = - \frac{\omega}{\rho g} = 1.7 \text{ cm sec}^{-1}.$$

The computed values are close to those observed in pre-trough situations in the earth's atmosphere.

In the earth's atmosphere, clouds do not usually extend above the tropopause. Let us assume that the mean height of the tropopause is at the jet stream level, a reasonable assumption. The jet stream level is characterized

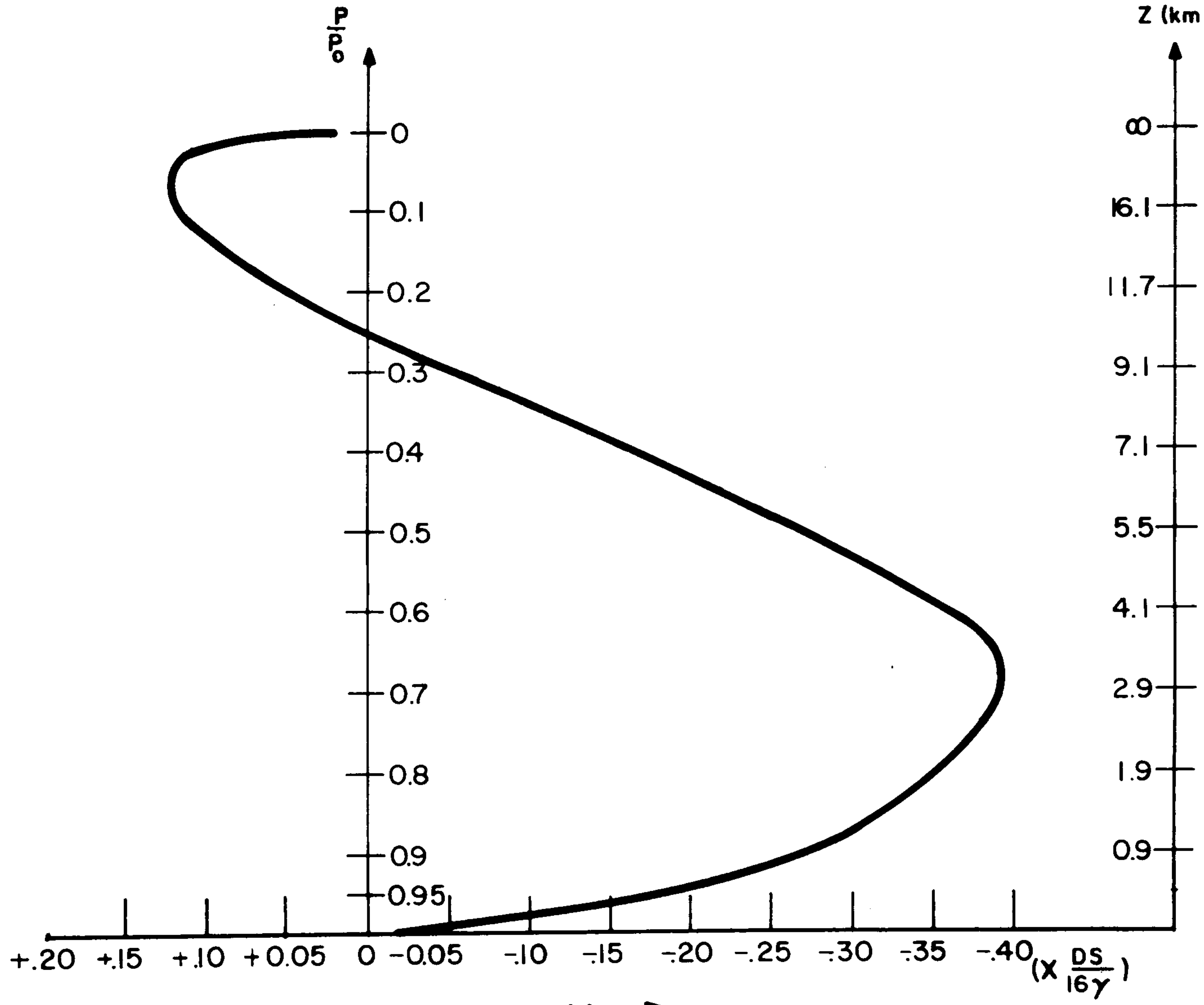

Figure 7. Computed vertical velocity profile for terrestrial atmosphere ($u = v_o = 40$ m sec^{-1}; $L = 4000$ km; $|S| = 12$ m^2 mb^{-1} sec^{-1} deg^{-1}).

by a maximum of horizontal velocity divergence. From the equation of continuity, one can obtain the result that $(\partial^2\omega)/(\partial p^2) = 0$ at a level of maximum velocity divergence. Thus, working in reverse, one can estimate the height of the mean tropopause, and, hence, the mean maximum height of frontal clouds, from an estimate of the height at which $(\partial^2\omega)/(\partial p^2) = 0$. From the earth's atmosphere computed profile we obtain $(\partial^2\omega)/(\partial p^2) = 0$ at about 350 mb, in reasonable agreement with the observed tropopause pressure of 250 mb at middle latitudes.

For disturbances on Mars, we may assume that

$$u_o = v_o \approx 60 \text{ m sec}^{-1}.$$

$$|S| = 90 \text{ m}^2 \text{ mb}^{-1} \text{ sec}^{-2} \text{ deg}^{-1}.$$

$$L = 4000 \text{ km}.$$

$$\rho = 4 \times 10^{-5} \text{ g cm}^{-3}.$$

The results are shown in Figure 8. The maximum vertical velocity is approximately

$$w = 12.8 \text{ cm sec}^{-1}$$

which is about six times larger than that computed for earth. Based on a 20 km scale height for the Martian atmosphere, $(\partial^2\omega)/(\partial p^2) = 0$, and, hence the mean tropopause is at 20 km. Thus, observations of Martian frontal clouds at a height of 30 km (See Figure 5 in Section 1) above the surface in lower latitudes, where one expects a higher than average tropopause, are probably correct. If frontal clouds can reach 30 km, the convective clouds on Mars can probably reach greater altitudes.

The height of the Martian tropopause computed on the basis of the dynamical considerations discussed above should be compared with estimates of the height of the Martian tropopause obtained from other considerations. Estimates of Martian tropopause height based upon a theoretical calculation of the vertical distribution of Martian temperature yield a value of 9 km [26]. However, the computed temperature profile [26] suggests that the Martian temperatures still decrease with height above the tropopause, albeit at a rate less than in the Martian troposphere. In the earth's atmosphere, temperatures actually increase with height above the tropopause. These differences in vertical temperature structure suggest that the Martian stratosphere may not have as strong a damping effect on vertical motions as the earth's stratosphere. Hence, dynamical systems may extend well above the tropopause level as defined by the vertical temperature distribution. This may explain the difference between the heights of the computed dynamic tropopause and the height computed by Ohring [26].

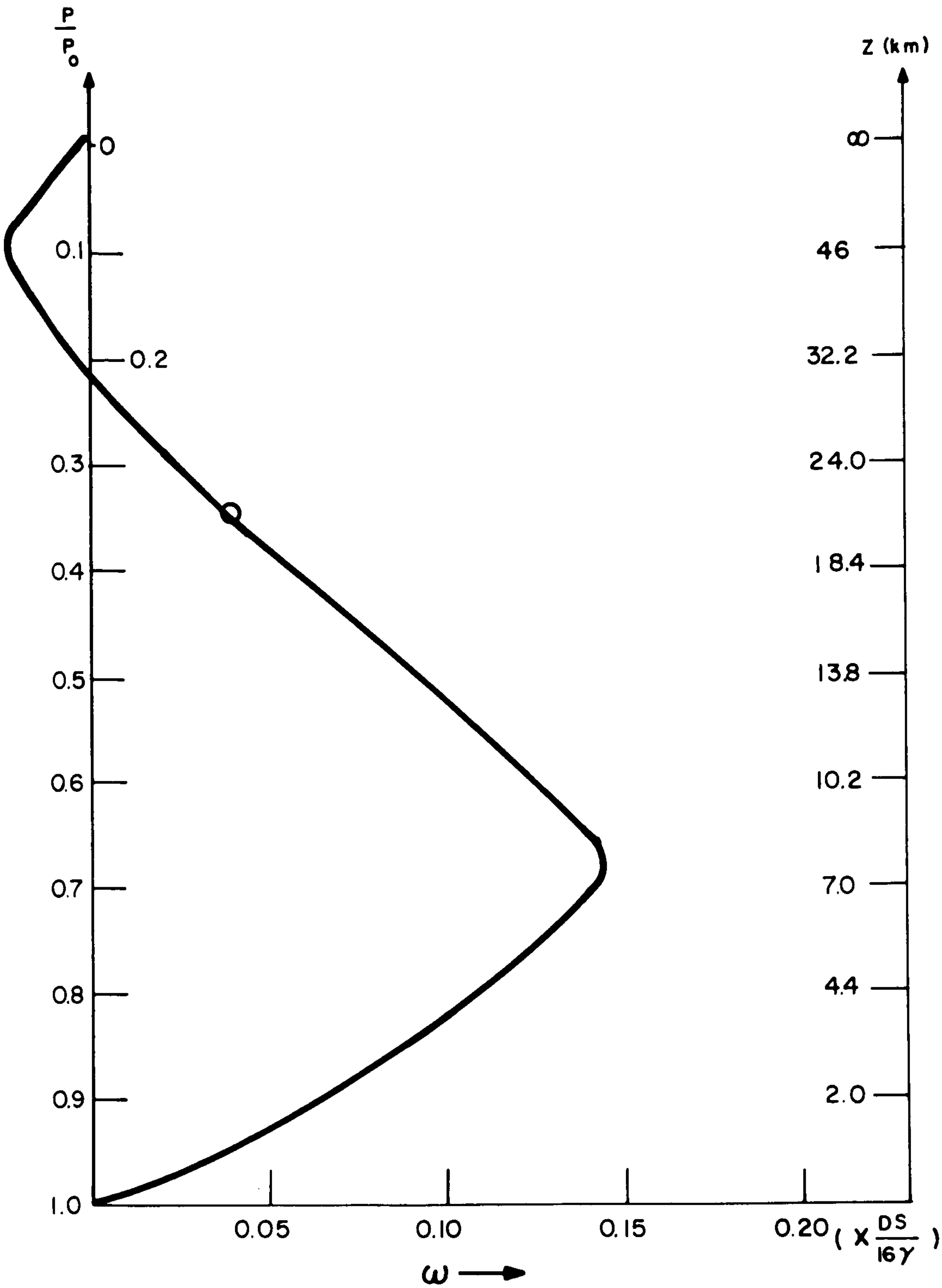

Figure 8. Computed vertical velocity profile for Martian atmosphere (u_o=v_o=60 m sec^{-1}; L=4000 km; $|S|$ = 90 m^2 mb^{-1} sec^{-1} deg^{-1}).

6. CONCLUSIONS

Analysis and interpretation of some Martian cloud observations leads to certain inferences about Martian circulation characteristics. A tendency for dust cloud systems to form at low latitudes of the winter hemisphere is noted. The trajectories of these cloud systems suggest the presence of a wave type circulation regime, upper level meridional flow at equatorial latitudes, and subtropical high pressure areas in the Martian atmosphere. Some observations of the vertical development of clouds suggest the possibility of fronts in the equatorial regions of Mars.

Two theoretical criteria are used to determine whether the Martian atmospheric circulation is of the wave type or of the symmetrical type. The application of the thermal Rossby number to the Martian atmosphere suggests a wave type circulation regime. The use of Mintz's [2] criterion, with an appropriate value of the dynamic coefficient of eddy viscosity, also predicts a wave type regime.

Based on Haurwitz's [21] circulation model, the average zonal and meridional winds are computed to be about 25 m sec^{-1} and 1.3 m sec^{-1}, respectively, for a surface pressure of 25 mb; and about 15 m sec^{-1} and 0.7 m sec^{-1} for a surface pressure of 85 mb. These values are higher than the average winds in the earth's atmosphere.

Theoretical estimates of the maximum surface winds on Mars suggest a value of more than 100 m sec^{-1}. The estimates are based upon Kuo's [24] theory for the maximum wind in a convective vortex, and an application of the gradient wind equation to a cloud system — assumed to represent a storm — on Mars.

From a simplified form of the "ω" equation, theoretical estimates of the vertical velocity profile in a Martian atmospheric wave are obtained. These calculations indicate that the maximum large scale vertical velocity of a disturbance in a baroclinic atmosphere is about 13 cm sec^{-1}. The mean tropopause height — i.e., a "dynamic" tropopause based upon the vertical velocity computations rather than a "radiative" tropopause based upon the vertical temperature structure — is about 20 km, in agreement with observations of the heights of frontal type cloud tops.

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