Mechanisms for Mars Dust Storms

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ABSTRACT

Characteristics of the Mars global dust storm are reviewed. At the Mariner 9 encounter, the dust consisted of highly absorbing particles distributed rather uniformly up to great height (~50 km). These observations together with temperature distributions inferred from the Mariner 9 IRIS by Hanel and his collaborators are used to estimate global wind systems during the dust storm. The global distribution and direction of light surface streaks indicate that the axially symmetric circulation was a dominant part of flow during the dust storm. An energy balance argument is used to estimate the intensity of the equatorial part of such a wind system. Surface winds driven by the diurnal and semi-diurnal components of solar heating are also estimated. The axially symmetric winds may become strong enough to raise dust over wide areas of Mars' tropics under unusual conditions: the incoming solar radiation must be near its seasonal maximum, the static stability must be low, and the atmosphere must be able to absorb and re-emit a sizeable fraction of the incoming radiation. Emission is aided by the formation of H2O ice clouds in the winter northern polar region. Absorption is enhanced by the presence of a small dust opacity in the atmosphere prior to the onset of the global dust storm. Strong winds around the periphery of the retreating south polar cap would be driven by the temperature gradient at the cap edge and by the mass outflow due to subliming CO2. These polar winds could generate local dust storms, raising the general level of dustiness, and providing the conditions necessary for onset of a global dust storm. Observational evidence for this sequence of events is discussed. The proposed model is sensitive to the precise phase relationship between Mars' perihelion and southern summer solstice, and variations in this phasing may have caused a strongly episodic behavior of dust storms and of a number of related planetary processes.

1. Introduction

The Mariner 9 spacecraft observed Mars during one of the most intense and long-lived dust storms ever seen. The resulting data are relevant to questions of dust storm properties, origin and life history, even though the initial phase of the storm was not observed by Mariner 9. These observations as well as ground-based observations have stimulated a number of efforts to understand the origin and decay of the dust storm (Golitsyn, 1973; Hess, 1972; Gierasch and Goody, 1973). In this paper we first review the evidence on dust storm properties, then attempt to develop a model of the surface wind distribution during the main phase of the storm. Only the simplest components of the global winds are discussed: the mean meridional circulation, and the diurnal and semi-diurnal wind systems. Other features, such as topographic wind systems, were undoubtedly of great importance, but may modify the main results only in detail.

After reviewing the factors which appear to have been important in driving these global wind systems during the storm, we are led to postulate a model for global dust storm generation. The model involves absorption of solar radiation in an atmosphere already containing a small concentration of dust, confined mainly to the Southern Hemisphere. During late spring in the Southern Hemisphere, the temperature difference between the retreating CO2 cap and the adjacent uncovered ground averages 90°C, and the resulting baroclinic zone coupled with the mass outflow current from the subliming cap provide reasonable mechanisms for generating local storms which could supply the low dust concentrations. Observational evidence is reviewed for both the presence of dust in the atmosphere prior to the main storm and for strong winds near the cap edge. The mechanisms proposed here for dust storm generation depend on close phasing between Mars' southern summer solstice and perihelion. They are thus sensitive to variations in Mars orbital parameters.

2. Observed characteristics of the global dust storm of 1971

The beginning phase of the dust storm has been described by Capen and Martin (1972), Baum (1973),
and Abramenko and Naugolnaya (1972). The major activity began 22 September in the eastern part of Noachis. The area covered by dust spread rather slowly within the Southern Hemisphere at first, but within a little over two weeks the entire planet was obscured. The location and seasonal date of origin of the storm corresponded closely with the time and place of origin of the 1956 storm. On 22 September, Mars was 8° past its perihelion; at the beginning of the 1956 storm, Mars was 2° past perihelion (Fig. 1). Although the major dust storm activity began on 22 September according to Capen and Martin, Abramenko and Naugolnaya have reported dust storm precursors as early as the beginning of September and yellow-red clouds in the Noachis-Hellasputus and Aerona-Arabia regions between 12 and 16 September. CO₂ pressure observations by Parkinson and Hunten (1973) also indicate the presence of some dustiness in the Hellas area as early as August. After the initiation of the main phase of the dust storm, dust spread gradually westward, but new centers of dust storm activity soon developed at other longitudes in the same latitude belt (Capen and Martin, 1972; Parkinson and Hunten, 1973).

During the first month after the arrival of Mariner 9 (13 November–13 December), the dust appeared to be almost uniformly distributed horizontally with an optical depth of about 2 (Masursky et al., 1972). Profiles of brightness on the limb and terminator showed that the dust was also mixed fairly uniformly in height up to at least 50 km (Leovy et al., 1972; Ajello et al., 1973). The color and albedo of the dust-shrouded planet were similar to those of the light surface areas indicating that the single-scattering albedo of the dust particles was low; the value \( \alpha_n \approx 0.8 \) near 600 nm has been estimated (Leovy et al., 1972). This low atmospheric brightness and the inferred low value of \( \alpha_n \) are consistent with characteristic particle size of order 5–10 μm. Th correlation between the ultraviolet spectrometer (UVS) and television data supports this general size estimation and also suggests a narrow range of sizes; both instruments measured comparable optical depths and cle Arlington during the dust storm (Hord et al., 1972; Leovy et al., 1972). If the preponderant particle size had been less than 1 μm, the UVS would have seen larger optical depths and slower clearing rates than the TV cameras. The Mariner 9 infrared interferometer spectrometer (IRIS) data also indicated predominant particle sizes in the 1–10 μm range (Conrath et al., 1973).

The IRIS data also show absorption and emission features due to silicate materials in the atmosphere, and both IRIS and radio occultation data showed that the dust strongly heated the atmosphere (Hanel et al., 1972; Kliore et al., 1972). According to the IRIS data, the heating effect of the dust extended up to at least 40 km, supporting the evidence from the UVS and the TV pictures that the dust was mixed to extraordinarily high levels. During this period, a persistent haze layer was observed at a height of about 70 km. This layer disappeared as the dust storm dissipated. It is likely that the high haze was composed of CO₂ ice condensed at a high-level temperature minimum (Briggs, 1972, private communication). An active convective layer is expected above the heated dusty part of the atmosphere, or above the level at which thermal contact between the dust particles and the air is lost, and a temperature minimum would occur at the top of the convective layer. If this interpretation is correct, the high condensate layer occurring with the dust storm is further
indirect evidence that the atmosphere was strongly heated up to the 40–50 km level.

The IRIS data have been analyzed in terms of averages over latitude and local time for the storm period (Fig. 2, from Hanel et al., 1972). The correspondence between the temperature variations at 2 mb (~10 km) and 0.3 mb (~30 km) indicates that the diurnal heating was nearly independent of height in the Southern Hemisphere.

Fig. 3 shows the density-weighted average atmospheric temperature obtained by combining the data of Fig. 2 with a meridional cross section of temperature obtained by Hanel and his collaborators from data on Mariner 9 revolution 102 (24 December). We believe that this single cross section is representative of the late phase of the dust storm because of the close relationship between the region of maximum horizontal temperature gradient and the southern edge of the north polar hood, which, in turn, was consistently found in the 45°–50° latitude zone by the TV (Leovy et al., 1972). For comparison, the average temperature from the general circulation model calculations of Leovy and Mintz (1969) for the corresponding season is shown in Fig. 3.

The higher observed temperatures at high southern and northern latitudes may be due to the absorption of heat by the dust and enhanced northward heat transport during the dust storm.

These observed characteristics of the dust storm will be used in the following discussion of global wind systems during the storm. To summarize: during the 13 November–13 December period, the dust storm could be characterized by optical depth ~2, nearly uniform mixing ratio of dust, and strongly absorbing particles (ratio of absorption to extinction ≥20%) causing heating which was nearly uniform in height up to at least 40 km.

3. Global winds during the dust storm

The temperature distributions shown in Fig. 2 can be used to infer characteristics of three components of the global wind system: the steady axially symmetric component, and the diurnal and semi-diurnal tides. Aspects of these wind systems have been discussed by Hanel et al. (1972) and by Conrath et al. (1973). The axially symmetric wind in the tropics is of particular interest since the global distribution of light albedo markings trailing from craters and other topographic obstacles clearly indicates a predominantly axially symmetric surface flow toward the latitude belt 25–30°S (the latitude of the subsolar point during the season of the dust storm). This streak pattern shows the influence of the Coriolis force: streaks extend from NE to SW in the northern tropics and from NW to SE in the southern tropics (Sagan et al., 1973). Since we are concerned here primarily with the intensity of this circulation near the equator, we use a very simple model of the global heat balance to estimate this intensity. The heat balance is used because the alternative, use of the momentum balance constraint, is particularly difficult near the equator; temperature gradients and momentum damping must be known with great precision. We then use the same heat balance model to estimate the magnitude of the wind and temperature variations in the thermal tides. Since the diurnal temperature oscillation is known, this approach provides a check on the simplified heat balance model which we use.

![Fig. 2](https://example.com/diagram2.png)

**Fig. 2.** Diurnal and latitude temperature variations, averaged over the first 42 days of the Mariner 9 orbiting period at the 2- and 0.3-mb levels. [Reproduced from Hanel et al. (1972).]

![Fig. 3](https://example.com/diagram3.png)

**Fig. 3.** Temperatures averaged in the vertical (weighted with density), and averaged over longitude and time of day. Mariner 9 data are from Hanel et al. (1972) [heavy solid line], and general circulation calculations from Leovy and Mintz (1969) [thin solid line]. Dashed portions of the Mariner 9 curve are interpolated or extrapolated.

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a. Global heat balance and axially symmetric winds in the tropics

The temperature data suggest that, below about 5 scale heights, the density $\rho$ at height $z$ can be approximated by

$$\rho \approx \rho_s e^{-z/H},$$

(1)

where $\rho_s$ is the mean density at the surface and $H = 10$ km is the scale height. Since variations of temperature horizontally and with local time are nearly independent of height (Fig. 2), we can approximate the dependence of the zonally averaged temperature on latitude $\phi$ and on height $z$ by

$$\bar{T}(\phi,z) \approx \Theta(z) \theta(\phi),$$

(2)

where $\Theta(z)$ is a slowly varying function of height. To good accuracy, the mean zonal west-to-east wind, $\bar{u}(\phi,z)$, is related to the zonally averaged temperature by the thermal wind equation:

$$\bar{u}(\phi,z) = \bar{u}_s(\phi) - \frac{g}{a f} \int_0^z \frac{1}{\bar{T}} \frac{\partial \bar{T}}{\partial \phi} \, dz,$$

(3)

where $\bar{u}_s(\phi)$ is the mean zonal surface wind, $a$ is the planetary radius, $f(\phi) = 2\Omega \sin\phi$ for rotation rate $\Omega$, and $g$ is the acceleration of gravity. Since $\Theta(z)$ is nearly independent of height, it follows from (1) and (3) that

$$u(\phi) \approx U(\phi) - \bar{u}(\phi)(1 - z/H),$$

(4)

where

$$U(\phi) = (H \rho_s)^{-1} \int_0^\infty \rho \, dz,$$

(5)

and $\bar{u}(\phi)$ is a height-independent wind shear. The south-to-north meridional wind, $\bar{v}(\phi,z)$, must have a similar dependence on height:

$$\bar{v}(\phi,z) \approx V(\phi) - \bar{v}(\phi)(1 - z/H).$$

(6)

Of course, Eqs. (4) and (6) cannot hold through the entire atmosphere, but they may be expected to hold approximately through the heated region, up to 4–5 scale heights. This is high enough that neglect of departures from (4) and (6) at high levels will not seriously affect the results. If the meridional mass flux vanishes,

$$V(\phi) = (\rho_s H)^{-1} \int_0^\infty \rho \, dz = 0,$$

(7)

and the surface meridional wind is

$$\bar{v}_s(\phi) = -\bar{v}(\phi).$$

(8)

The steady, axially symmetric heat balance can be written

$$\frac{1}{a \cos\phi} \frac{\partial}{\partial \phi} \left[ \cos\phi \int_0^\infty \rho \bar{u} \left( C_p \bar{P} + g \bar{z} \right) \, dz \right] = \Delta F,$$

(9)

where $C_p$ is the specific heat at constant pressure, and $\Delta F$ is the net heating per unit area of an atmospheric column (Lorenz, 1967). Substituting (1), (2), (6) and (8) into (9), we obtain an expression for the surface meridional wind needed to maintain the energy balance:

$$\frac{1}{a \cos\phi} \frac{\partial}{\partial \phi} \left[ \bar{u}_s(\phi) \cos\phi \right]$$

$$= g \Delta F \left( C_p \bar{P} \left[ \kappa + H T^{-1} \left( \partial T/\partial \phi \right) \right] \right)^{-1},$$

(10)

where $P_s$ is the globally averaged surface pressure, $\kappa = R / C_p$ with $R$ the gas constant for CO$_2$, and $\partial T/\partial \phi$ the density-weighted global average of $\partial T/\partial \phi$. The factor $\left[ \kappa + H T^{-1} \left( \partial T/\partial \phi \right) \right]$ is proportional to the static stability. During the part of the dust storm observed by Mariner 9, this factor was about 0.13; reduction of the static stability leads to a corresponding increase in the intensity of the axially symmetric circulation.

In order to estimate the net heating, we make the following assumptions:

(i) The atmosphere loses heat, primarily to space, at temperatures corresponding to those in Fig. 3, and with uniform emissivity $\varepsilon$.

(ii) The atmosphere absorbs a constant fraction $\alpha$ of the insolation at each latitude. Because the atmosphere is stably stratified during the storm, convective heat input is neglected.

The assumptions of uniform absorptivity and emissivity are based on the premise that uniformly distributed dust is responsible for absorption, and both dust and CO$_2$ are responsible for emission. The approximation that emission takes place at the average temperature shown in Fig. 3 is based in part on the observation that the vertical variation of temperature during the dust storm was not large and in part on the radiative properties of a dusty CO$_2$ atmosphere; radiation takes place effectively from all levels (see below). Temperature differences between the atmosphere and ground are much smaller during the dust storm than normally (Hanel et al., 1972), so the atmosphere-ground exchange has been omitted in calculating the global energy balance. Because there is considerable diurnal cancella-
tion of this exchange rate, it should be much smaller than the loss of energy to space. The ratio $\alpha/\varepsilon$ is determined by the requirement that the net heating of the atmosphere as a whole vanishes. We obtain $\alpha/\varepsilon=0.69$, and the heat input and loss factors shown in Fig. 4; their difference is the net heating $\Delta F$. The energy flux scale for Fig. 4 is fixed by our assumed emissivity, $\varepsilon=0.35$. This is based on an emissivity of $\sim0.17$ for CO$_2$ at Mars atmosphere temperatures, and a comparable emissivity for dust, crudely estimated from the spectra of Hanel et al. The combined effect of CO$_2$ and dust is to radiate energy effectively from all levels of the atmosphere. The CO$_2$ tends to radiate most efficiently from levels near 30 km. The dust on the other hand, with an optical depth of order unity in the infrared, tends to radiate efficiently from all levels. If there is any tendency for size sorting of the particles with height, the dust will actually radiate primarily from the lower part of the atmosphere, so that both lower and upper portions of the atmosphere cool efficiently to space (Sargent and Beckman, 1973).

An alternative pair of heat input and loss curves is obtained by taking into account the effect of H$_2$O ice clouds. Only 1.0 mg cm$^{-2}$ of ice in clouds under Martian conditions will produce an emissivity of $\sim0.7$ (Leovy, 1966). Allowing for additional emissivity due to CO$_2$ and dust, and taking into account the distribution of the north polar hood clouds as observed by the TV cameras, we estimate a value of $\varepsilon=0.8$ north of 47 N. The resulting energy balance factors are also shown in Fig. 4. The role of the ice cloud emission is to produce a small change in the value of $\alpha$ required for balance; $\alpha$ increases from 0.235 to 0.258. Despite the small change in $\alpha$, when the requirement for global energy balance is retained, the ice cloud emissivity produces a large change in the net heating $\Delta F$ at all latitudes. The ice clouds form a very efficient radiator which must be supplied with heat by the large-scale atmospheric circulation.

Fig. 5 shows the meridional surface wind (equal to the negative of the meridional wind shear $\theta$), calculated from Eq. (10) and the condition $\theta=0$ at the poles. It can be seen that the effect of the radiating clouds is to double the strength of the meridional circulation required to maintain the heat balance, from a maximum value of $\sim9$ m sec$^{-1}$ to a maximum of $\sim20$ m sec$^{-1}$. The meridional circulation shown is unlikely to be valid outside of the equatorial zone because eddy transport of heat is probably dominant there. In the general circulation model calculations, the mean meridional circulation carried most of the energy between $\pm25^\circ$ latitudes (Leovy and Mintz, 1969). This is the same zone in which there is a systematic, axially symmetric distribution of streaks. Also shown in Fig. 5 is the average vertical wind shear, obtained from Eq. (3) and the temperature distribution of Fig. 3.

The vertical shears of the meridional and zonal winds, $\theta$ and $\varphi$, can be used to estimate the magnitude of the zonal component of the surface wind. If the contribution to momentum flux by quasi-horizontal eddies is neglected, the equation of motion for the mean zonal flow is

$$ \frac{\partial \bar{U}}{\partial t} = \frac{1}{\rho} \frac{\partial}{\partial x} \left[ \bar{U} \frac{\partial \bar{U}}{\partial x} + \bar{V} \frac{\partial \bar{V}}{\partial y} + \bar{W} \frac{\partial \bar{W}}{\partial z} \right], $$

where $\bar{W}$ is the mean vertical velocity, and $\bar{U}$ the zonal component of the vertical stress (positive for upward flux of zonal momentum). An equation for $\partial U/\partial t$ is obtained by multiplying (11) by $\rho$, integrating over height, and making use of (4), (6), and the continuity equation

$$ \frac{\partial}{\partial x} \left[ \frac{1}{\rho} \frac{\partial \bar{U}}{\partial x} \right] - \rho \frac{\partial \bar{U}}{\partial y} = 0. $$

The result is

$$ \frac{\partial U}{\partial t} = \frac{1}{\rho} \frac{\partial}{\partial x} \left[ \bar{U} \frac{\partial \bar{U}}{\partial x} + \bar{V} \frac{\partial \bar{V}}{\partial y} + \bar{W} \frac{\partial \bar{W}}{\partial z} \right] - \rho \frac{\partial \bar{U}}{\partial y}, $$

expressed in terms of the friction velocity $\bar{u}_f$. For steady-state conditions and vanishing total mass flux ($V=0$), Eq. (13) reduces to

$$ \bar{u}_f = \rho \frac{\partial}{\partial x} \left[ \frac{\partial \bar{U}}{\partial x} + \frac{\partial \bar{U}}{\partial y} + \frac{\partial \bar{U}}{\partial z} \right]. $$

The zonal surface stress balances the meridional transport of zonal momentum. The approximate value of the friction velocity derived from (14) and (15) and the $\theta$ and $\varphi$ distributions of Fig. 5 is shown in Fig. 6, with the factor $\bar{u}_f/(\bar{u}_f^2+\bar{v}_f^2)^{1/2}$ arbitrarily set equal to unity. The convention used in the figure is that $\bar{u}_f$ is
positive for surface winds from the west. The scale relationship between $\bar{u}_y$ and $\bar{u}_x$ shown in Fig. 6 is based on the empirical relation,

$$\frac{\bar{u}_y}{\bar{u}_x} \approx 0.05,$$

approximately valid under moderate or high wind conditions (Csanady, 1972).

b. Diurnal and semidiurnal thermal tides

We assume that the diurnally varying heating is uniform up to some height $h \ll H$, and vanishes above that level, and that the total heating in each diurnal harmonic is equal to the product of that harmonic component of the insolation and the fractional absorption $\alpha$. Given the global heating, the diurnal and semidiurnal wind fields can be calculated using the classical tidal theory as described by Chapman and Lindzen (1970). The heating is expanded in a series of Hough functions $\Theta^s_n$ which are the eigenfunctions of Laplace's tidal equation with frequency $\sigma$ (normalized by the rotational frequency so that $\sigma = 1$ for the diurnal tide), longitudinal wavenumber $s$, and latitudinal mode number $n$. Thus, given $(\sigma, s)$, the corresponding tidal wind components are

$$u^{s,\sigma} = \sum_n \left( U_n^{s,\sigma} V_n^{s,\sigma} \right) \left( \frac{d}{dx} \frac{y_n^{s,\sigma}}{2} y_n^{s,\sigma} \right). $$

The expansion functions $U_n^{s,\sigma}, V_n^{s,\sigma}$ are linear combinations of the corresponding Hough functions $\Theta^s_n$ and its derivative, and are graphed in Chapman and Lindzen for the more important modes. The vertical coordinate $x$ is defined by $dx/dz = H^{-1}$, which is nearly independent of height for Martian dust storm conditions. The vertical structure of the tidal wind and temperature components is given in terms of the solution $y_n^{s,\sigma}$ of

$$\frac{d^2 y_n^{s,\sigma}}{dx^2} + \nu x y_n^{s,\sigma} = \frac{g^2 H_n^{s,\sigma}}{4 \Omega^2 \alpha},$$

where the vertical structure parameter is

$$\beta^s_n = \left[ \frac{\nu H_n^{s,\sigma}}{4 \Omega^2 \alpha} \left( \frac{H}{\partial T} \right) \right]^{-1}.$$

The $\Lambda^s_n$ are eigenvalues of Laplace's tidal equation and can be calculated from values given by Chapman and Lindzen. The quantity $g(\Delta F_n^{s,\sigma})/P$ is the thermally-tidal heating per unit mass for a given mode $(\sigma, s, n)$.

We now assume that the heating (per unit mass) is uniform up to some height $h \ll H$, and vanishes above that level, and that the total heating in each harmonic is equal to the product of that harmonic component of the insolation and the fractional absorption $\alpha$. With these assumptions

$$(\Delta F_n^{s,\sigma}) = \left( \frac{\alpha H_n^{s,\sigma}}{1 - \alpha}, \right) \right), \quad 0 \leq x \leq x_0$$

where $Q_n^{s,\sigma}$ is the contribution of the $(\sigma, s, n)$ mode to the incoming solar radiation and $x_0$ is the depth of the heated layer. Utilizing (19), the vertical structure equation (17) can be solved analytically. The appropriate boundary conditions are

$$\frac{dy_n^{s,\sigma}}{dx} + \left( \frac{g^2 H_n^{s,\sigma}}{4 \Omega^2 \alpha} \right) y_n^{s,\sigma} = 0$$

at $x = 0$, and, for $x > x_0$,

$$y_n^{s,\sigma} \approx \exp \left[ -\beta^s_n |x| \right].$$

In this section we use the conventional tidal theory notation that $s^\sigma$ and $\sigma^s$ are the southerly and westerly wind components, respectively.

The eigenvalue $\Lambda_n^{s,\sigma} = 4 \Omega^2 \alpha / y_n^{s,\sigma}$, where $y_n^{s,\sigma}$ is the equivalent depth. The functions $\phi_n^{s,\sigma}$ defined by (17) correspond to those in Chapman and Lindzen divided by $y_n^{s,\sigma}$, where $\gamma$ is the ratio of specific heats.

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Fig. 6. Zonal surface wind from momentum balance, using values of $\bar{u}$ and $\bar{v}$ from Fig. 3 (with $\epsilon = 0.80$ north of 47N, and assuming $\bar{u}_{\alpha} = 0.05$. Positive values indicate wind from the west.

Fig. 7. Amplitude factor for the diurnal harmonic of the insolation (heavy solid curve), and its representation by a linear combination of the four lowest-order symmetric and four lowest-order antisymmetric Hough functions (thin solid curve). The dashed curve is the semidiurnal harmonic of the insolation. The ordinate gives the amplitudes as fractions of the solar constant.
when $\beta_n^s < 0$, as it may be for the diurnal tide, or
\[ y_n^s = \exp\left[i\beta_n^s \left| \frac{dx}{x} \right| \right] \] (22)
when $\beta_n^s > 0$, as it is for the semi-diurnal tide if $H(\partial T/\partial z) = -17K$. In evaluating the solutions to (17)-(22), we have assumed that $H(\partial T/\partial z) = -20K$, and have taken the limit $\exp(-a) \to 0$. The surface wind and temperature fluctuations are not sensitive to $\delta T/\delta z$ in the range of interest nor to the value of $x_0$ provided that $x_0 \geq 4$.

The diurnal harmonic of the insolation has been expressed in terms of the eight lowest-order diurnal Hough functions, four symmetric and four antisymmetric about the equator. The fit of this truncated expansion to the insolation is shown in Fig. 7. The resulting diurnal surface wind pattern is dominated by the $\Theta_1^1$ mode in the subtropics and the $\Theta_{23}^2$ mode in high latitudes. The surface pressure oscillation associated with the lowest asymmetric Hough function vanishes identically, so that the surface wind field does not show marked asymmetry about the equator. The maximum surface wind is about 20 m sec$^{-1}$ in the subtropics.

The semi-diurnal heating harmonic, also shown in Fig. 7, can be fitted very accurately with only the lowest symmetric semi-diurnal Hough function (see p. 126 of Chapman and Lindzen). The surface wind pattern associated with this function is highly divergent with outflow away from points near local noon and local midnight on the equator, and inflow toward equatorial points near 0600 and 1800 local time (Fig. 8). These winds are strongest in the tropics and subtropics, and have an amplitude of $\sim 8$ m sec$^{-1}$ at latitudes equatorward of $\pm 45^\circ$.

The diurnal temperature oscillation has been calculated using the same heating model and an expression analogous to (16), and the results have been compared with the diurnal temperature oscillation estimated.
from the data of Hanel et al. (1972). This comparison is shown in Fig. 9 for the density-weighted, vertically-averaged temperatures. It can be seen that the comparison of amplitudes in the Southern Hemisphere is satisfactory, for \( \alpha = 0.4 \). The discrepancy between the value of \( \alpha \) required to fit the diurnal temperature oscillation, and that estimated for global heat balance may be attributed to our highly simplified heating model, but two other factors may also contribute.

1) The diurnal harmonic may be overestimated in the observed temperature data. This could occur if there is a significant contribution to temperature variance along latitude circles by standing waves. Conrath (1972, personal communication) reports that there is a small contribution to temperature variance by standing topographic waves.

2) The emissivity and absorptivity may be underestimated. It is unlikely that the emissivity due to dust and \( \text{CO}_2 \) has been underestimated; if anything, our value of 0.35 may be somewhat high. But ice clouds at latitudes other than the north polar region could contribute substantially to the daily average emissivity. Nighttime temperatures fall to about 190 K over a wide latitude and height range. At this temperature, 1 mg cm\(^{-2}\) of water, uniformly mixed as vapor, would condense and form ice clouds. Ice clouds occurring only at night might enhance the average daily emissivity by as much as 50%. Nightside ice clouds could affect the atmospheric energy balance very differently during dust-storm conditions than during non-dust-storm conditions. During the dust storm, the entire atmospheric column can cool to condensation at night, and ice clouds can act to cool the entire atmospheric column even further. Normally ice clouds could only affect a shallow layer near the surface. In order to maintain global energy balance, the value of \( \alpha \) would have to be increased in proportion to the daily average emissivity increase, and the amplitudes of all the wind systems discussed above would also increase proportionately. There is some evidence that ice cloud formation was occurring on the night side during the dust storm. TV pictures taken on the dark side of the morning terminator on Mariner 9 revolution 50 showed surprising levels of illumination, nearly uniform spatially, at solar depression angles of 12° and 14°. The surface brightness was \( \approx 1/1000 \) of that of the illuminated side. No similar illumination was seen on the dark side of the evening terminator. A possible cause for this puzzling observation is multiple scattering by layers of ice clouds on the morning side.

The phase comparison between observed temperatures and those computed from the tidal theory is also shown in Fig. 9. The computed maximum occurs near 1800 local time, later than the observed maximum by \( \approx 1-2 \) hr. Part of this discrepancy may be due to the neglect of radiative damping in our application of the theory. Inclusion of this factor would shift the phase of the computed temperature maxima for both the diurnal and semi-diurnal components, and would also shift the wind patterns, causing them to occur earlier by as much as one hour. Such a shift is indicated in Fig. 8.

Despite these discrepancies, the gross agreement between the observed and computed diurnal temperature fields lends some support to the heat balance model used for calculating the axially symmetric circulation. The sense of the discrepancy in the diurnal temperature amplitude indicates that the fractional absorptivity \( \epsilon \) should be increased to about 0.40. This would imply a corresponding increase in the axially symmetric meridional wind near the equator to a maximum \( \mathbf{u} \) value near 30 m sec\(^{-1}\), and a maximum \( \mathbf{v} \) value near 25 m sec\(^{-1}\).

4. Mechanisms for global dust storms

a. A scenario for the dust storms

The dominant wind system deduced in the preceding section was the axially symmetric circulation. Zonally-averaged surface wind speeds in the equatorial zone may have been \( \approx 30 \) m sec\(^{-1}\) during this period. This is marginally strong enough to raise dust by saltation of larger particles (Golitsyn, 1973). The pattern of these winds in the equatorial zone corresponds well with the pattern of light surface streaks. On the other hand, the surface winds due to the diurnal and semi-diurnal tides were not strong enough to raise dust, although they contribute significantly to the total wind.

The factors responsible for generating intense meridional circulations are strong insolation, low static stability, and high atmospheric absorptivity and emissivity. More specifically, the enhanced emissivity due to ice clouds in the north polar hood zone was shown to have a strong influence on the circulation intensity even in equatorial latitudes.

With these factors in mind, the following sequence of events is proposed as a plausible scenario for development of global dust storms such as that of 1971. During the southern spring, local dust storms produced by one of several mechanisms to be discussed below occur mainly in the Southern Hemisphere. The dust content of southern, middle, and subtropical latitudes gradually builds up, perhaps to optical depths of a few tenths, and absorption of solar radiation takes place preferentially in these Southern Hemisphere latitudes. The optical depth and absorption rates continue to increase until perihelion. The emissivity remains relatively low (\( \epsilon \approx 20\% \)) and, coupled with low temperatures near the 30-km level of maximum \( \text{CO}_2 \) emission, the small emissivity keeps the outgoing radiation relatively low in the south. The direct rate of absorption of solar energy is much less at this time than during the main phase of the global dust storm, but this is partially compensated by convective heat flux from the surface, which may amount to \( \approx 8\% \) of the incoming solar
the diurnal tides also shift the surface by as much as several kilometers. According to the general circulation model calculations, the average static stability at this season is about half as large as that at high latitudes. Dust absorption is neglected as it was observed to be during the storm. From Eq. (10), this static stability decrease translates into a factor of 2 increase in the intensity of the meridional circulation required for energy balance. Thus, it is reasonable to expect the axially symmetric part of the equatorial surface winds to reach ~30 m sec⁻¹ prior to the onset of the storm.

The observations show that the initial development of the storm occurs in a particular longitude belt, that of Hellas Planitia. This initiation may be due to the superposition of one or more additional wind systems on top of a meridional circulation which is already strong. Because the northeastern part of Noachis is an especially favored area for storm generation, the topography of that region is probably essential as suggested by Gierasch and Goody (1973). The "dusty hurricane" suggested by Gierasch and Goody may also play a role. The diurnal and semidiurnal tides may also contribute although we expect them to be relatively weak at the beginning of the storm when there is less solar absorption. During the time of day when tides augment the axially symmetric and topographic components of the flow (near local noon), the boundary layer turbulence is most intense, and saltation can be most easily initiated.

Once dust begins to be raised on a very large scale in the southern subtropics, the intensity and depth of the wind system will continue to increase at first. As the circulation depth increases, the dust depth and depth of the heated layer increase correspondingly. Eventually the dust depth reaches several scale heights, the static stability increases as a result of the heating, and the meridional and topographic circulations, both of which vary inversely with the static stability, begin to weaken. The dust storm enters its decaying phase, the phase observed by Mariner 9. These aspects of the storm behavior are similar to the behavior of the "dusty hurricane model" investigated by Gierasch and Goody, but they occur for any circulation system driven by dust heating.

b. Generation of local dust storms

The dust storm postulated above depends on intense heating for its initiation. We have assumed that pre-existing dustiness, limited to small optical depths, is responsible for the heating. There are several mechanisms by which low levels of dustiness could be created and maintained in Southern Hemisphere latitudes. Among these are dust devils, and the local cyclonic dust storms investigated by Gierasch and Goody. Both of the latter would occur preferentially in the southern subtropics at Mars' perihelion season.

We call attention here to two other mechanisms which are likely to be important. The first is the intense baroclinic zone which must prevail near the edge of the
receding polar cap in the spring. General circulation model calculations have indicated that strong surface winds are likely on Mars in the most baroclinic regions (Leovy and Mintz, 1969), and Mariner 9 pictures have shown the development of a local dust storm in the baroclinic zone near the north polar hood (Leovy et al., 1972). As the spring season progresses, the temperature gradients at the edge of the cap would increase at first as the temperature of the bare ground adjacent to the CO₂ ice rises. As the cap shrinks, it eventually becomes less effective in cooling air which flows over the cap interior. Thus maximum temperature gradients would occur at some intermediate season. Although estimates are difficult, this maximum presumably occurs just before or during the rate of maximum shrinking of the south cap, when Mars is between heliocentric longitudes 300° and 330° (Fig. 10a). In regions where the surface slopes downward away from the cap, the winds may be analogous to the katabatic winds flowing off the Antarctic Continent, but with a stronger thermal drive.

An additional factor which is strongly seasonally dependent is the mass outflow produced by CO₂ sublimation over the cap. The longitudinally averaged outflow is not large, but it can produce a large Coriolis torque. From Eq. (13), the Coriolis torque due to a mass outflow V is proportional to fV. The latter is approximately balanced by the surface stress:

\[
fV = \frac{\tau_v}{\rho H} - \frac{u_a^2}{H},
\]

when non-steady effects, and nonlinear terms are neglected. The mass outflow wind can be computed by assuming that the excess radiation at the cap surface causes sublimation of CO₂; heat storage and transport of heat by the atmosphere are neglected. The net radiation depends on the emission temperature of the CO₂ surface, which is well known, and the albedo of the cap, which is not well known. The outflow wind is an increasing function of the ratio of cap area to cap circumference. Fig. 10b shows the computed outflow wind based on the seasonally varying insolation (Cross, 1971), and the observed latitude variation of the edge
of the cap (Michaux, 1967). Because of differences in the seasonal size variations of the north and south polar caps, and in the two radiation balances, the maximum value of the mass outflow wind from the north cap is only about half as large as that from the south cap; and the mechanism operates more effectively in the south than in the north. The relatively large mass outflow from the south cap reaches a maximum about 45 days before Mars’ perihelion (point S in Fig. 1) and falls off slowly after the maximum. Thus, the timing of the southern mass outflow is consistent with the idea that it may contribute to general dustiness in the southern hemisphere at perihelion. The actual flow of mass from the cap is unlikely to be uniform; in all probability, it is channeled into locally stronger outflow currents around the cap periphery.

c. Observational evidence for seasonal variations in the background level of dustiness and for polar winds

The model we have postulated attributes the main phase of the global dust storm to a meridional circulation driven by solar energy absorbed in already somewhat dusty air in the Southern Hemisphere. The strongest evidence for the intense meridional circulation during the main phase of the storm is the surface streak distribution in Mars’ tropics reported by Sagan et al. (1973). Fig. 11, reproduced from Masursky et al. (1972), shows streaks in the dusty air flowing southward from the “South Spot” caldera (105, 120W) during the storm. The direction is consistent with the postulated mean meridional flow near the surface, but it may be modified by the topographic flow which would be southward near the surface, and by the atmospheric tides which would be toward the south at this time (late morning) and latitude.

There is some evidence, apart from the global dust storms themselves, that there is seasonal variation in the background level of dustiness, with the maximum dustiness during the southern spring and early summer. Spacecraft data (Mariners 4, 6, 7 and 9) have shown Mars’ atmosphere to be hazier during southern spring and summer than during northern spring and summer (Leovy et al., 1973). Fig. 12 shows the distribution of haze layers observed by Mariners 6 and 7 during early spring in the south (point 6 in Fig. 1). There is a tendency for the layers to occur preferentially in the middle southern latitudes, and color filter data at point P in Fig. 12 show the haze at that point to be dust-colored material rather than condensates. Mars’ atmosphere may have been quite hazy prior to the onset of the 1956 storm. Polarization is a sensitive indicator of haziness; it decreases at all phase angles from “normal” values for a “clear” Mars. According to Dolkhus (1965), the polarization was anomalously small when he first began to observe the planet during July of 1956 (heliocentric longitude η of ∼300°). The major storm did not begin until late August of that year (η=356°). During 1971, the contrast of classical surface markings decreased a few weeks prior to the onset of the main storm (Baum, 1973). Although other interpretations are possible, this decrease may have been due to an increase in the background level of dustiness.

There is also considerable evidence that very strong surface winds occur near the periphery of the polar caps during the seasons in which CO₂ sublimation is occurring. Sagan et al. (1973) have shown that the pattern of dark markings on the floors of craters is related to the pattern of streaks extending from craters. Since the streaks relate to wind directions at a particular time (a time of exceptionally strong winds), the dark markings are also related to the wind directions at these times. In fact, there is good evidence from the Mariner 9 data that the explanation advanced by Sagan and Pollack (1969) for the seasonal changes in surface albedo is correct, i.e., that the changes are due to removal and restoration of light mobile dust from dark areas (Sagan et al., 1972; Leovy et al., 1972). Cutts (1973) has shown that the locations and orientations of dark markings around the periphery of the south polar cap indicate that the markings were shaped by southeasterly flow off the cap, and that erosional features in the south polar region also were formed by flow from the southeast. The features are most evident near 70S, the latitude at which the maximum rate of cap recession occurs (Fig. 10). The observed features were presumably formed by the strongest winds, and the southeasterly flow is that expected during the spring season of cap recession, both for the sublimation flow itself and for the wind component driven by the local temperature gradients. Cloud patterns near the edge of the north cap during spring also indicate winds with an easterly component (Leovy et al., 1973). To the extent that the so-called “wave of darkening” can be explained by surface albedo variations, the behavior of this phenomenon also supports the model. Pollack et al. (1967) have analyzed the time-dependent behavior of the “wave,” and have shown a general tendency for darkening to begin in the spring in each hemisphere. They show that darkening occurs earliest at the highest latitudes, at least in a statistical sense. This behavior is at least consistent with the idea that light material is removed from the surface by winds preferentially during the spring, although other interpretations are also possible.

Two TV frames from late in the Mariner 9 mission support this view of polar winds. Fig. 13 shows the north cap in early summer. The prominent dark band around the cap may well be due to removal of light material from the surface by winds flowing out from the cap as it recedes. Fig. 14 shows a curious atmospheric feature which has been described elsewhere (Leovy et al., 1973): bands of haze extend outward from the cap along the evening terminator, and recurve toward the west near latitude 45°. It is possible that
these streaks result from winds flowing off the cap, confined to a relatively narrow jet. Although these two figures show the north cap, we have seen that the maximum outflow winds from the south cap are likely to be stronger than those in the north.

5. Discussion

The global dust storm of 1971 was not an isolated event. Similar large storms have been seen near most perihelic oppositions (Golitsyn, 1973). The model developed here attempts to isolate factors which may be responsible for this regular recurrence.

The model contains a number of speculative details, but the factors which enter the model and which favor dust storm development are less speculative. These are: 1) proximity of Mars to the sun; 2) existence of effective atmospheric radiators, either in the winter hemisphere, or perhaps both in the winter hemisphere and on the night side; and 3) capacity for the atmosphere to absorb a sizeable fraction of the insolation. We have shown that there are both theoretical and observational reasons to expect that springtime katabatic and mass outflow winds from the subliming south polar cap supply a seasonally varying background level of dust in the atmosphere, preferentially in the Southern Hemisphere, which is capable of providing the required capacity for absorption.

Departures from axial symmetry must be important at the beginning of the global dust storms. Topographic
factors seem to favor storm origin in Noachis, and the non-uniform dust distribution during the early stages must cause major circulation asymmetries during the initial growth of the dust cloud. But the surface streak distribution points to the predominance of the axially symmetric circulation following this initial phase. The distribution of energy sources and sinks needed to drive such a storm occurs only near the solstices, and the apparent restriction of these storms to perihelion indicates that the combination of insolation and atmospheric absorptivity and emissivity needed to produce a global dust storm occurs only at the southern summer (perihelion) solstice. Thus, we may expect variations in the frequency of global dust storms associated both with the eccentricity of Mars' orbit and with the phasing of the perihelion and solstices. Long-period cycles in the orbital eccentricity, and in the phase relation between perihelion and solstices, have been discussed by Murray et al. (1973).

Thus, as Briggs (1973) has also pointed out, the frequency of global dust storms is likely to vary markedly with variations in the astronomical processes, and we may expect similar variations in erosion and deposition processes. Cold storage, removal and transport of volatiles is also sensitive to the occurrence of global dust storms. The data of Hanel et al. and Kilore et al. have shown that the 1971 dust storm profoundly altered atmospheric temperatures, and we have shown that the rate of meridional heat transport was also affected. The meridional heat transport has a strong influence on the rate at which volatiles can condense in the winter polar caps and on the H₂O/CO₂ ratio in the condensate (Leovy, 1973). Thus, periodic variations in the frequency of occurrence of the global dust storms provide a plausible mechanism for forming the laminated structures found in the polar regions (Murray et al., 1972). The possible link between the global dust storms, the polar laminae, and long-period variations in the seasonal distribution of insolation at Mars has been explored in more detail by Briggs.

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