Chapter 8 Discussion

8.1 Possible Feedback between Saltation Threshold and Dust Storm Activity

It was shown that the interannual variability of the GDSs can be reproduced by the LOM with stochastic forcing for a range of parameter values. Thus, the results of the LOM simulations are consistent with the hypotheses of extrinsic irregularity. The frequency and intensity of the storms depend on the values of the threshold friction speed, the amplitude of the dust source, the standard deviation of the stochastic process and the particle size distribution on the surface. The time and place of occurrence of the GDSs and accompanying changes in the wind and temperature fields are consistent with the observations. However, only one GDS can occur during a year in the LOM, while observations suggest that there are years with two dust storms.

Noise added to the LOM may represent random variations in the atmospheric heating due to variations in albedo, thermal inertia, passage of a weather system, and so on, or variations in the value of the threshold friction speed arising from aeolian activity.

The response of the LOM to noise is similar in the cases when noise is added to the temperature fields or to the threshold friction speed. However, noise added to the amplitude of the surface dust source does not result in the interannual variability.

The occurrence of interannual variability in the LOM is very sensitive to the value of the threshold friction speed. Changes of the order of several per cent in the values of the threshold friction speed can lead to cessation of the interannual variability in the occurrence of GDSs in the LOM. For the randomly varying threshold friction speed the allowed variability is larger, but is still of the order of tens of per cent.

Such a strong dependence on one parameter is quite suspicious. This could happen if there is a feedback between the amplitude of the dust source s_0 or the standard deviation of the random process and the threshold friction speed. If these parameters decrease and increase together with the threshold friction speed, then the interannual variability is possible in a wide range of parameters. For example, for the case shown in Fig. 7.2, if the value of U_t^* is lowered from 1.19 to 1.17, the number of storms increases to 11 in 15 years, and for $U_t^* = 1.16$ the GDSs occur every year. However, if the amplitude of the dust source s_0 is decreased to a sufficiently smaller value, the interannual variability is still possible even for $U_t^* = 1.16$. Yet, such a dependence is hard to justify. Another possibility involves a feedback between the value of the threshold friction speed and atmospheric dust activity.

One can envision the following mechanism for the proposed feedback: increased atmospheric activity and stronger winds lead to erosion and increase in roughness of the surface, by exposing buried rocks, pebbles or other non-erodible particles. This in turn leads to an increase of the threshold friction speed, due to aerodynamic sheltering of the erodible particles by non-erodible roughness elements [GI85, p. 82-85]. On the other hand, decreased atmospheric activity favors accumulation of sand (dust) and decrease in surface roughness, leading to a decrease in the value of the threshold friction speed. [GI85] give the following approximate relationship between the rough surface threshold speed and roughness, derived from the wind tunnel experiments of [LSS74]:

$$U_{tR}^*/U_t^* = 2(D_p/z_0)^{-1/5} \tag{8.1}$$

where the subscript, R, refers to rough surface threshold, D_p is particle diameter and z_0 is equivalent roughness height. The equivalent roughness height z_0 is the measure of the surface roughness [GI85, GLLT92]. It is a function of the roughness element's shape, Reynolds number and the distance between the elements. For a quiescent sand surface, z_0 is $\sim 1/30$ the sand particle diameter.

If the system is started with a low threshold friction speed, the GDSs follow each year. Strong winds during the GDSs erode the surface and remove saltating particles so that the roughness of the surface and the threshold friction speed both increase. The removed particles accumulate in the opposite hemisphere or at sites where GDS initiation is inhibited. Observations by Pathfinder indicate that repeated burial and exhumation of rocks has taken place at the landing site in the past [GB00]. The threshold friction speed continues to increase until it reaches the critical value, at which point interannual variability is possible.

After that, if there were no stabilizing process resupplying particles to eroding sites from the sites of accumulation, the increase in threshold friction speed pushes the system into the state without dust storms. On the other hand, if there is a stabilizing process, like dust devils or local winds resupplying particles to the GDS generating sites, a balance between erosion and resupply of particles may be reached. If the system has a high threshold friction speed, there will be no GDSs and material will build up, again supplied by dust devils or other mechanisms, until small rocks are buried and the roughness of the surface is decreased. The threshold friction speed decreases as well, which gives rise to the GDSs. In this case, the threshold wind speed will always be close to the critical value for occurrence of GDSs.

Appendix A Dust Transport Equation

Let us consider derivation of the dust transport Eq. (2.3) for the domain a of the atmosphere (see Fig. 2.4). The goal is to express continuity Eq. (2.22) in terms of average dust optical depths of the amospheric domains a, b, c and d (Eq. (2.23)). For simplicity, we will assume that $p_t = 0$. Multiplying Eq. (2.22) by $\sigma^* \cos\theta$ and integrating over domain a yields:

$$\frac{\partial \tau_a}{\partial t} = -\frac{\tau_a}{t_d} + \frac{g\sigma^*}{2\pi r_m}\left(\iint_a \frac{\partial \psi}{\partial p}\frac{\partial q}{\partial y}dpdy - \iint_a \frac{\partial \psi}{\partial y}\frac{\partial q}{\partial p}dpdy\right) \tag{A.1}$$

where the use of (2.8)-(2.9) was made. Integrating by part yields:

$$\frac{\partial \tau_a}{\partial t} = -\frac{\tau_a}{t_d} + \frac{g\sigma^*}{2\pi r_m}\left(\int_{p_s/2}^{0}\left(q\frac{\partial \psi}{\partial p}\right)_{y=0} dp - \int_{-L}^{0}\left(q\frac{\partial \psi}{\partial p}\right)_{p=p_s/2} dy\right) \tag{A.2}$$

Equation (A.2) states that change of opical depth in the domain a is due to dust fluxes through boundaries between a and b (first integral in Eq. (A.2)) and a and d (second integral in Eq. (A.2)). Assuming that dust is distributed approximately uniformly within a given domain, τ can be expressed in terms of q as:

$$\tau \approx \sigma^* \frac{p_s r_m}{2\sqrt{2}} q \tag{A.3}$$

where τ and q represent average optical depth and dust mixing ratio of a given domain.

It can be shown that the asymmetric part of the stream function $\bar{\psi}_a$ does not contribute to dust transport to domain a. Indeed, the flow assosiated with $\bar{\psi}_a$ does not cross the vertical boundary between a and b, and the net transport through the horizontal boundary between a and d is zero. Thus, ψ in Eq. (A.2) can be substituted by $\bar{\psi}_s$:

$$\frac{\partial \tau_a}{\partial t} = -\frac{\tau_a}{t_d} + \frac{g\sigma^*}{2\pi r_m}\left(\left(q\bar{\psi}_s\right)_{y=0} - \left(q\bar{\psi}_s\right)_{p=p_s/2}\right) \tag{A.4}$$

The direction of the fluxes on the boundaries depends on the sign of ψ_s. For positive ψ_s (air rises in NH) the flux through boundary between a and b is from b to a and the flux through boundary between a and d is from a to d. For negative ψ_s the direction is reversed. Hence we arrive at the dust transport equation for domain a:

$$\frac{\partial \tau_a}{\partial t} = -\frac{\tau_a}{t_d} + a\bar{\psi}_s(\tau_a - \tau_d)(1-\delta) + a\bar{\psi}_s(\tau_b - \tau_a)\delta \tag{A.5}$$

where

$$\delta = \begin{cases} 0, & \psi_s \geq 0 \\ 1, & \psi_s < 0, \end{cases}$$

and $a = g\sqrt{2}/p_s r_m^2 \pi$.

Transport equations for domains b, c and d are derived in an analogous way.

Appendix B Heating Rate and Seasonal Changes

The basis of the approach is the assumption that the atmosphere can be divided into regions (a, b, c, d in Fig. 2.4) for which equilibrium temperatures depend only on the amount of dust in the region and seasonal date. The data from [HLP82] allows one to determine the functional dependence of the equilibrium atmospheric temperatures in different regions of the atmosphere on the amount of the atmospheric dust. The dependence of the radiative damping time t_r on dust optical depth is approximated based on the net heating calculations from [PHSL90]. The seasonal dependence is obtained from a simple radiative model, discussed below. The desired equilibrium temperatures for nonuniform dust distributions and different seasons are then calculated as a product of the function describing the dependence on the atmospheric dust amount and the seasonally varying equilibrium temperatures for clear atmosphere.

First, the dependence on dust is discussed. The atmosphere was divided into four domains, as before (see Fig. 2.4). It was assumed that for each domain one can determine an average characteristic equilibrium temperature, denoted by T_a^e, T_b^e, T_c^e and T_d^e on Fig. 2.4. The LOM's equilibrium temperatures can be expressed using the just introduced characteristic equilibrium temperatures in the following way:

$$T_{av}^e = \left(T_a^e + T_b^e + T_c^e + T_d^e\right)/4 \tag{B.1}$$

$$T_{ns}^e = -\left(T_a^e - T_b^e - T_c^e + T_d^e\right)/4k \tag{B.2}$$

$$T_v^e = \left(T_a^e + T_b^e - T_c^e - T_d^e\right)/4k \tag{B.3}$$

where $k = 2/\pi$. The above expressions follow from the definition of the average characteristic equilibrium temperatures in the four boxes model:

$$
\begin{aligned}
T_a^e \equiv \frac{2}{Lp_s}\iint_A T^e dpdy &= T_{av}^e - kT_{ns}^e + kT_v^e && \text{(B.4)}\\
T_b^e &= T_{av}^e + kT_{ns}^e + kT_v^e && \text{(B.5)}\\
T_c^e &= T_{av}^e + kT_{ns}^e - kT_v^e && \text{(B.6)}\\
T_d^e &= T_{av}^e - kT_{ns}^e - kT_v^e && \text{(B.7)}
\end{aligned}
$$

Due to its particular functional form T_{em}^e does not enter into Eqs. (B.1)-(B.3). Hence, it was assumed that it changes in the same way as T_{av}^e, since both of these functions are constant with height and symmetric relative to equator.

It was further assumed that the average temperatures in these four domains depend on the season and the average dust optical depth in the domains in the following way:

$$
\begin{aligned}
T_a^e(\tau_a) &= T_a^e(0)f_a(\tau_a) && \text{(B.8)}\\
T_b^e(\tau_b) &= T_b^e(0)f_b(\tau_b) && \text{(B.9)}\\
T_c^e(\tau_c) &= T_c^e(0)f_c(\tau_c) && \text{(B.10)}\\
T_d^e(\tau_d) &= T_d^e(0)f_d(\tau_d) && \text{(B.11)}
\end{aligned}
$$

where $T_i^e(0)$ is the equilibrium temperature of the ith domain for clear condition (i being, as before, one of the subscripts a, b, c or d). $T_i^e(0)$ is seasonally varying, while $f_i(\tau_i)$ is varying both with dust and season (see below). The functions $f_i(\tau_i)$ were assumed to have the following form:

$$
f_i(\tau_i) = \frac{(1 - q_i e^{-s_i \tau_i})}{1 - q_i} \qquad \text{(B.12)}
$$

where the coefficients q_i and s_i were determind from [HLP82] by fitting the values of the average characteristic temperatures T_i^e for given τ_i. Coefficients q_i and s_i are given in Table 2.1. The functions $f_i(\tau_i)$ are equal to 1 when $\tau_i = 0$, and tend to a value of $1/(1 - q_i)$ as optical depth increases.

The functions $f_i(\tau_i)$ mimic the behavior of the temperature fields in the GCM [PHSL90]. The atmospheric temperatures are very sensitive to addition of even small amounts of dust to the clear atmosphere. For example, the GCM experiments with $\tau = 0.3$ show a significant increase in atmospheric temperatures compared to the case $\tau = 0$. However, for a dusty atmosphere the effect of addition of dust is less pronounced. In the GCM experiment the temperature fields with $\tau > 1$ show little difference (for the same season). Accordingly, the function $f_i(\tau_i)$ increases sharply from $\tau_i = 0$ to $\tau_i = 1$ and then "levels off" for $\tau_i > 1$.

Analysis of the GCM results [HLP82, PHSL90] suggests that dust is thermodynamically less "active" in the lower part of the winter atmosphere due to the lower solar angle and increased cooling during the longer nights that offsets heating during the day. Accordingly, the relationship between "active" dust and season is:

$$\tilde{\tau_c} = \tau_c(1 + J)/2 \quad \text{(B.13)}$$

$$\tilde{\tau_d} = \tau_d(1 - J)/2 \quad \text{(B.14)}$$

where $J = \theta_0/\theta_{0max}$, θ_0 is the subsolar point latitude, $\theta_{0max} = 25°$ and J varies from 1 at northern summer to −1 at southern summer. τ_c and τ_d are the actual dust optical depths in the lower atmosphere above the NH and SH, respectively, and $\tilde{\tau_c}$ and $\tilde{\tau_d}$ are the corresponding "active" dust optical depth that are to be substituted into Eq. (B.12).

Finally, to determine the heating rate, the dependence of the radiative damping time t_r on dust optical depth is needed. The dependence can be roughly approximated from the Ames GCM experiments ([PHSL90], Fig. 4, 5). In general, the effect of dust is to shorten the radiative damping time, but the effect levels off for high dust amounts. For simplicity, it was assumed that the radiative damping time is spatially constant throughout the atmosphere. [PHSL90] present calculations of the net solar heating Q for $L_s = 279°$ and for dust optical depth varying from 0 to 5. From Eq. (2.6)

$$t_r = \tilde{t_r} = -\frac{T - T^e}{Q}$$

As a proxy for $T - T^e$, we use the value of $T_{av} - T^e_{av}$, which is obtained from 2D GCM calculations [HLP82] of temperature fields and radiative-convective temperature fields for the same conditions. The resulting dependence was modeled as

$$\tilde{t}_r = \tilde{t}_{0r}/f_r(\tau_r) \tag{B.15}$$

where $\tilde{t}_{0r}$ is the value of the radiative damping time at $L_s \approx 270°$ for a clear atmosphere and is a parameter of the model. f_r has the same structure as f_i, given by Eq. (B.12), and τ_r is the average optical depth:

$$\tau_r = (\tau_n + \tau_s)/2 = (\tau_a + \tau_b + \tau_c + \tau_d)/2$$

where τ_n and τ_s are the total optical depths in the NH and SH respectively. The coefficients q_r and s_r of the function f_r were determind by fitting the values of $(T - T^e)/Q$ for given τ_r. The values of q_r and s_r are given in the Table 2.1.

To calculate the seasonal dependence of the equilibrium temperatures $T^e_i(0)$ of the clear atmosphere, a simple radiative model was used. It was assumed that seasonal changes in insolation translate into seasonal changes in the maximum atmospheric equilibrium temperatures. As the seasons progress from southern summer to northern summer, the maximum of the atmospheric equilibrium temperature moves from the SH to the NH. In the process, T^e_{ns} switches sign. Accordingly, the maximum equilibrium temperature T^e_{max} of the atmosphere during southern or northern summer is

$$T^e_{max} = T^e_{av} + |T^e_{ns}| - T^e_v$$

(since $T^e_{ns} < 0$ during southern summer). Due to the eccentricity of the martian orbit, the subsolar insolation at perihelion ($L_s = 251°$, southern spring) is 43% larger than at aphelion. Assuming that the solar heating Q is balanced by infrared cooling proportional to $\sigma(T^e_{max})^4$, where σ is the Stefan-Boltzman constant, it follows that

$$T^e_{max} \sim Q^{1/4}$$

and T^e_{max} changes by about 11% seasonally. Assuming that the changes are spread equally between the components of the T^e_{max} (T^e_{av}, $|T^e_{ns}|$, T^e_v), the change of every component between southern and northern summer is about 11% by absolute value. The exact form of the seasonal change law is derived from calculations of the change in solar heating Q using Kepler's law. In addition, the T^e_{ns} term changes sign as the subsolar point switches between southern and northern hemispheres. This change was assumed to be proportional to θ_0/θ_{0max} and is superimposed on the 11% change between southern and northern summers.

The equilibrium temperatures $T^e_i(0)$ are then calculated using Eqs. (B.4)-(B.7).

Appendix C Low-Order Model

The equations of the reduced model are written in term of dimensionless variables:

$$X_1 = u_a/U, \quad X_5 = T_{av}/T, \quad X_9 = \tau_a/\tau,$$
$$X_2 = u_s/U, \quad X_6 = T_{ns}/T, \quad X_{10} = \tau_b/\tau,$$
$$X_3 = \psi_a/\Psi, \quad X_7 = T_v/T, \quad X_{11} = \tau_c/\tau,$$
$$X_4 = \psi_s/\Psi, \quad X_8 = T_{em}/T, \quad X_{12} = \tau_d/\tau,$$

where the normalization coefficients are:

$$\begin{aligned} U &= \frac{\Omega r_m}{2} \qquad (\mathrm{m}\, s^{-1}) \\ \Psi &= \frac{p_s r_m^2}{g t_d} \qquad (\mathrm{kg}\, s^{-1}) \\ T &= \left(\frac{H}{h}\right) \frac{r_m^2}{R t_d^2} \qquad (\mathrm{K}) \\ \tau &= 1 \qquad (\text{optical depth}) \end{aligned}$$

Here H is atmospheric scale height (~ 11 km) and h is the height of the model domain (~ 45 km). The equations of the LOM are:

$$\dot{X}_1 = -c_1 X_1 + a_{11} X_2 X_4 - a_{12} X_4 - a_{13} X_1 X_3 \tag{C.1}$$

$$\dot{X}_2 = -c_1 X_2 + a_{21} X_2 X_3 - a_{22} X_3 + a_{23} X_1 X_4 \tag{C.2}$$

$$\dot{X}_3 = -c_1 X_3 + a_{31} X_2 + a_{32} X_2^2 - a_{33} X_8 \tag{C.3}$$

$$\dot{X}_4 = -c_1 X_4 + a_{41} X_1 + a_{42} X_1 X_2 + a_{43} X_6 \tag{C.4}$$

$$\dot{X}_5 = -c_2 (X_5 - F_1) \tag{C.5}$$

$$\dot{X}_6 = -c_2 (X_6 - F_2) - a_{61} X_4 (X_7 + c_4 X_5 + a_{62} X_8) + a_{63} X_3 X_6 \tag{C.6}$$

$$\dot{X}_7 = -c_2 (X_7 - F_3) + a_{71} X_4 X_6 - a_{72} X_3 X_7 \tag{C.7}$$

$$\dot{X}_8 = -c_2 (X_8 - F_4) + a_{81} X_3 (X_7 + c_4 X_5 + a_{82} X_8) + a_{83} X_4 X_6 \tag{C.8}$$

$$\dot{X}_9 = -X_9 + c_3 X_4 \left((X_9 - X_{12})(1 - \delta) + (X_{10} - X_9)\delta\right) \tag{C.9}$$

$$\dot{X_{10}} = -X_{10} + c_3X_4\left((X_{10} - X_9)(1-\delta) + (X_{11} - X_{10})\delta\right) \tag{C.10}$$

$$\dot{X_{11}} = -X_{11} + X_{10} + c_3X_4\left((X_{11} - X_{10})(1-\delta) + (X_{12} - X_{11})\delta\right) + + s^*\delta + s_n^* \tag{C.11}$$

$$\dot{X_{12}} = -X_{12} + X_9 + c_3X_4\left((X_{12} - X_{11})(1-\delta) + (X_9 - X_{12})\delta\right) + + s^*(1-\delta) + s_s^* \tag{C.12}$$

where a dot denotes a derivative with respect to the dimensionless time $t' = t/t_d$, s^* is dimensionless interactive dust source described in Section 2.5, s_s^* and s_s^* are dimensionless independent dust sources in the SH and NH, and

$$\delta = \begin{cases} 0, & X_4 \geq 0 \\ 1, & X_4 < 0. \end{cases}$$

The circulation and dust are linked via forcing functions F_1, F_2, F_3 and F_4, which are the normalized equilibrium temperatures:

$$F_1 = T_{av}^e/T, \quad F_2 = T_{ns}^e/T, \quad F_3 = T_v^e/T, \quad F_4 = T_{em}^e/T.$$

The method, by which the dependence of the forcing functions on dust was estimated, is described in Section 2.4.

The coefficients of the model are given in Table C.1:

Table C.1: Coefficients of the LOM.

a_{11}	0.12	a_{12}	0.37	a_{13}	0.2
a_{21}	0.38	a_{22}	2.0	a_{23}	0.52
a_{31}	$0.053(\Omega t_d)^2$	a_{32}	$0.016(\Omega t_d)^2$	a_{33}	0.78
a_{41}	$0.44(\Omega t_d)^2$	a_{42}	$0.17(\Omega t_d)^2$	a_{43}	1.9
a_{61}	0.65	a_{62}	0.1	a_{63}	0.18
a_{71}	0.52	a_{72}	0.45		
a_{81}	0.43	a_{82}	0.3	a_{83}	0.06
c_1	rt_d	c_2	t_d/t_r	c_3	0.19
c_4	0.24				

Bibliography

[AGM+83] R. E. Arvidson, E. A. Guinness, H. J. Moore, J. Tillman, and S. D. Wall. Three Mars years: Viking Lander 1 imaging observations. *Science*, 22:463–468, 1983.

[AHL85] S. G. Ahler, P. C. Hohenberg, and M. Lucke. Thermal-convection under external modulation of the driving force .1. The Lorenz model. *Phys. Rev. A*, 32:3493–3518, 1985.

[AL78] E. M. Anderson and C. B. Leovy. Mariner 9 television limb observations of dust and ice hazes on Mars. *J. Atmos. Sci.*, 35:723–734, 1978.

[Bag41] R. A. Bagnold. *The physics of blown sand and desert dunes.* Methuen, New York, 1941.

[Con75] B. J. Conrath. Thermal structure of the Martian atmosphere during the dissipation of the dust storm of 1971. *Icarus*, 24:36–46, 1975.

[CPH89] D. S. Colburn, J. B. Pollack, and R. M. Haberle. Diurnal variations in optical depth at Mars. *Icarus*, 79:159–189, 1989.

[Fow89] T. B. Fowler. Application of stochastic-control techniques to chaotic nonlinear-systems. *IEEE Trans. on Aut. Control*, 34:201–205, 1989.

[FPM97] L. K. Fenton, J. C. Pearl, and T. Z. Martin. Mapping Mariner 9 dust opacities. *Icarus*, 130:115–124, 1997.

[Fra78] K. Fraedrich. Structural and stochastic analysis of a zero-dimensional climatic system. *Quart. J. Roy. Met. Soc.*, 104:461–474, 1978.

[FZ95] M. Franz and M. Zhang. Supression and creation of chaos in a periodicaly forced Lorenz system. *Phys. Rev. E*, 52(4):3558–3565, 1995.

[GB00] M. P. Golombek and N. T. Bridges. Erosion rates on Mars and implications for climate change: Constraints from the Pathfinder landing site. *J. Geophys. Res.*, 105(E1):1841–1853, 2000.

[GG73] P. J. Gierash and R. M. Goody. A model of a martian great dust storm. *J. Atm. Sci.*, 30:169–179, 1973.

[GI85] R. Greeley and J. D. Iversen. *Wind as a geological process on Earth, Mars, Venus and Titan.* Cambridge Univ. Press, New York, 1985.

[GLLT92] R. Greeley, N. Lancaster, S. Lee, and P. Thomas. Martian aeolian processes, sediments, and features. In H. Kieffer, B. Jakosky, C. Snyder, and M. Matthews, editors, *Mars*, chapter 22. Univ. of Arizona Press, Tucson, 1992.

[GWP+76] R. Greeley, B. R. White, J. B. Pollack, J. D. Iversen, and R. Leach. Saltation threshold on Mars: the effect of interparticle force, surface roughness and low atmospheric density. *Icarus*, 29(3):381–393, 1976.

[Hab86] R. M. Haberle. Interannual variabilty of global dust storms on Mars. *Science*, 234:459–461, 1986.

[Has76] K. Hasselman. Stochastic climate models. Part I. Theory. *Tellus*, 28:473–485, 1976.

[HFT95] F. Hourdin, F. Forget, and O. Talagrand. The sensitivity of the Martian surface pressure and atmospheric mass budget to various parameters: A comparison between numerical simulations and Viking observations. *J. Geophys. Res.*, 100:5501–5523, 1995.

[HH80] I. M. Held and A. Y. Hou. Nonlinear axially symmetric circulation in a nearly inviscid atmosphere. *J. Atmos. Sci.*, 37:515–533, 1980.

[HHBY97] R. M. Haberle, H. Houben, J. R. Barnes, and R. E. Young. A simplified three-dimensional model for Martian climate studies. *J. Geophys. Res.*, 102(E4):9051–9067, 1997.

[HLP82] R. M. Haberle, C. B. Leovy, and J. B. Pollack. Some effects of global dust storms on the atmospheric circulation of Mars. *Icarus*, 50:322–367, 1982.

[Hol92] J. R. Holton. *An introduction to dynamic meteorology.* Academic Press, San Diego, 3rd edition, 1992.

[Hou86] J. T. Houghton. *The physics of atmospheres.* Cambridge University Press, Cambridge, 2nd edition, 1986.

[HPB+93] R. M. Haberle, J. B. Pollack, J. R. Barnes, R. W. Zurek, C. B. Leovy, J. R. Murphy, H. Lee, and J. Schaeffer. Mars atmospheric dynamics as simulated by the NASA Ames general circulation model. *J. Geophys. Res.*, 98:3093–3123, 1993.

[HRT+80] S. L. Hess, J. A. Ryan, J. E. Tillman, R. M. Henry, and C. B. Leovy. The annual cycle of pressure on Mars measured by Viking Landers 1 and 2. *Geophys. Res. Lett.*, 7:197–200, 1980.

[IL93] A. P. Ingersoll and J. R. Lyons. Mars dust storms: interannual variability and chaos. *J. Geophys. Res.*, 98(E6):10,951–10,961, 1993.

[IW82] J. D. Iversen and B. R. White. Saltation threshold on Earth, Mars and Venus. *Sedimentology*, 29:111–119, 1982.

[KMZL92] R. A. Kahn, T. Z. Martin, R. W. Zurek, and S. W. Lee. The martian dust cycle. In H. Kieffer, B. Jakosky, C. Snyder, and M. Matthews, editors, *Mars*, chapter 29. Univ. of Arizona Press, Tucson, 1992.

[LBY+72] C. B. Leovy, G. A. Briggs, A. T. Young, B. A. Smith, J. B. Pollack, E. N. Shipley, and R. L. Wildey. The Martian atmosphere: Mariner 9 television experiment progress report. *Icarus*, 17:373–393, 1972.

[LHS+79] G. F. Lindal, H. B. Hotz, D. N. Sweetnam, Z. Shippony, J. P. Brenkle, G. V. Harstell, R. T. Spear, and W. H. Michael Jr. Viking radio occultation measurements of the atmosphere and topography of Mars: Data acquired during 1 Martian year of tracking. *J. Geophys. Res.*, 84:8443–8456, 1979.

[Lor63] E. N. Lorenz. Deterministic nonperiodic flow. *J. Atmos. Sci.*, 20:130–141, 1963.

[LSS74] L. Lyles, R. L. Schrandt, and N. F. Schneidler. How aerodynamic roughness elements control sand movement. *Transactions, Amer. Soc. Agric. Eng.*, 17:134–139, 1974.

[LTGB85] C. B. Leovy, J. E. Tillman, W. R. Guest, and J. Barnes. Interannual varibility of Martian weather. In G. E. Hunt, editor, *Recent Advances in Planetary Meteorology*, pages 69–84. Cambridge University Press, New York, 1985.

[LZP73] C. B. Leovy, R. W. Zurek, and J. B. Pollack. Mechanisms for Mars dust storms. *J. Atm. Sci.*, 30:749–762, 1973.

[Mar74] L. J. Martin. The major martian yellow dust storm of 1971. *Icarus*, 22:175–188, 1974.

[Mar84] L. J. Martin. Clearing the martian air: The troubled history of dust storms. *Icarus*, 57:317–321, 1984.

[MHTP93] J. R. Murphy, R. M. Haberle, O. B. Toon, and J. B. Pollack. Martian global dust storms: zonally symmetric numerical simulations including size-dependant particle transport. *J. Geophys. Res.*, 98:3197–3220, 1993.

[MI80] S. Moriyama and T. Iwashima. A spectral model of the atmospheric general circulation on Mars: a numerical experiment including the effects of the suspended dust and topography. *J. Geophys. Res.*, 85:2847–2860, 1980.

[MPH+95] J. R. Murphy, J. B. Pollack, R. M. Haberle, C. B. Leovy, O. B. Toon, and J. Schaeffer. Three-dimensional numerical simulation of Martian global dust storm. *J. Geophys. Res.*, 100(E12):26,357–26,376, 1995.

[MSS99] J. A. Magalhaes, J. T. Schofield, and A. Seiff. Results of the Mars Pathfinder atmospheric structure investigation. *J. Geophys. Res.*, 104(E4):8943–8955, 1999.

[PCF+79] J. B. Pollack, D. S. Colburn, F. M. Flasar, R. Kahn, C. E. Carlston, and D. Pidek. Properties and effects of dust particles suspended in the Martian atmosphere. *J. Geophys. Res.*, 84(B6):2929–2945, 1979.

[Pet81] R. A. Peterfreund. Visual and infrared observations of wind streaks on Mars. *Icarus*, 45:447–467, 1981.

[Pet85] R. A. Peterfreund. *Contemporary aeolian processes on Mars: local dust storms.* PhD thesis, Arizona State Univ., 1985.

[PHSL90] J. B. Pollack, R. M. Haberle, J. Schaeffer, and H. Lee. Simulations of the general circulation of the Martian atmosphere. 1. Polar processes. *J. Geophys. Res.*, 95(B2):1447–1473, 1990.

[RSL81] J. A. Ryan, R. D. Sharman, and R. D. Lucich. Local Mars dust storm generation mechanism. *Geophys. Res. Lett.*, 8:899–902, 1981.

[Sch55] H. Schlichting. *Boundary-Layer Theory.* McGraw-Hill, New York, 1955.

[Sch83] E. K. Schneider. Martian great dust storms: interpretive axially symmetric models. *Icarus*, 55:302–331, 1983.

[SK77] A. Seiff and D. B. Kirk. Structure of the atmosphere of Mars in summer in mid-latitudes. *J. Geophys. Res.*, 82:4364–4378, 1977.

[SL99] P. H. Smith and M. Lemmon. Opacity of the martian atmosphere measured by the imager for Mars Pathfinder. *J. Geophys. Res.*, 104(E4):8975–8985, 1999.

[SM80] B. Saltzman and R. E. Moritz. A time dependent climatic feedback system involving sea-ice extent, ocean temperature and CO_2. *Tellus*, 32:93–118, 1980.

[Spa82] C. Sparrow. *The Lorenz equations: bifurcations, chaos, and strange attractors.* Springer, New York, 1982.

[SRF93] Y. Shao, M. R. Raupach, and P. A. Findlater. Effect of saltation bombardment on the entrainment of dust by wind. *J. Geophys. Res.*, 98:12,719–12,726, 1993.

[SWB91] J. Singer, Y. Z. Wang, and H. H. Bau. Controlling a chaotic system. *Phys. Rev. Lett.*, 66:1123–1125, 1991.

[Tho79] T. E. Thorpe. A history of Mars atmospheric opacity in the southern hemisphere during the Viking extended mission. *J. Geophys. Res.*, 84:6663–6683, 1979.

[Til85] J. E. Tillman. Martian meteorology and dust storms from Viking observations. In C. P. McKay, editor, *Case for Mars II.* Univelt, San Diego, 1985.

[Til88] J. E. Tillman. Mars global atmospheric oscillations: annually synchronized, transient normal mode oscillations and the triggering of global dust storms. *J. Geophys. Res.*, 93:9433–9451, 1988.

[TSCJ94] E. Tziperman, L. Stone, M. A. Cane, and H. Jarosh. El Niño chaos: overlaping resonances between the seasonal cycle and the Pacific Ocean-atmospherer oscillator. *Science*, 264:72–74, 1994.

[Whi79] B. R. White. Soil transport by winds on Mars. *J. Geophys. Res.*, 84:4643–4651, 1979.

[ZBH+92] R. W. Zurek, J. R. Barnes, R. M. Haberle, J. B. Pollack, J. E. Tillman, and C. B. Leovy. Dynamics of the atmosphere of Mars. In H. Kieffer,

B. Jakosky, C. Snyder, and M. Matthews, editors, *Mars*, chapter 26. Univ. of Arizona Press, Tucson, 1992.

[ZH89] R. W. Zurek and R. M. Haberle. Martian great dust storm: aperiodic phenomena? In S. Lee, editor, *MECA Workshop on Dust on Mars III*, pages 40–41. LPI Tech. Rept. 89-01, 1989.

[ZM93] R. W. Zurek and L. J. Martin. Interannual variability of planet-encircling dust storms on Mars. *J. Geophys. Res.*, 98:3247–3259, 1993.

[Zur82] R. W. Zurek. Martian great dust storms: An update. *Icarus*, 50:288–310, 1982.