Review Article

Influence of dust on the dynamics of the martian atmosphere above the first scale height

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Abstract

Dust suspended in the martian atmosphere strongly affects the radiative transfer. Diabatic heating and cooling it creates are prominent factors that drive the atmosphere at various scales. This paper provides a review of dust influence on the large-scale dynamics in the atmosphere of Mars above approximately 10 km. We outline the established properties of dust that influence the diabatic heating/cooling rates, and summarize the current knowledge of dust-related effects on the zonal-mean circulation and zonally asymmetric disturbances: planetary waves and tides.

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1. Introduction

Airborne dust suspended in the dry atmosphere of Mars plays a role in many ways similar to the role of water on Earth. Being strongly radiatively active, dust absorbs solar radiation and emits in the infrared. The created local heating and cooling affects the atmospheric dynamics at various scales: from synoptic and meso-scale weather systems (martian meteorology) to the large-scale circulation and climate. A “dust cycle” on Mars with lifting from the surface, transport, and sedimentation back to reservoirs resembles the “water cycle” on Earth. Many martian weather...
phenomena are associated with dust: storms from local to regional and global ones, dust devils (Balme and Greeley, 2006), “cells” (Cantor et al., 2002), clouds (Clancy et al., 2003), and plumes (Fuerstenau, 2006). These similarities with the terrestrial hydrometeorology prompted Heavens et al. (2011b) to introduce the term “coniometeorology” (from the Greek word konios for “dust”) to characterize the martian weather.

This paper is a review of the current knowledge of dust effects on the global-scale atmospheric dynamics of Mars. Most of meteorological phenomena are associated with the lower atmosphere where the influence of the surface is strong. In the dynamical meteorology of Earth, the definition of the “lower” atmosphere is vague, and most often means the troposphere. The martian troposphere, an atmospheric layer with vertically decreasing temperature, extends much higher, to 40–60 km. The martian middle atmosphere lies above, and covers the altitudes to 110–130 km. In this survey, the emphasis is placed on the layer which is little directly affected from below by convection, and, from above, by heating due to absorption of extreme ultraviolet and energetic particles. It corresponds to the altitudes between ~10 and 110–130 km, that is, coincides with the troposphere above approximately one scale height and the middle atmosphere of Mars. Therefore, we leave out of scope a large body of studies, which concern mechanisms of dust lifting and interactions with synoptic-scale disturbances, and focus on effects of the dust already injected in the air.

The most spectacular phenomena are the dust storms that regularly reach planetary scales. They sometimes obscure almost all the surface of the planet. Earlier observations of these storms on Mars started in the beginning of the 20th century with ground-based telescopes (Briggs et al., 1979, and references therein). Since the 1970s, landers and spacecraft inserted into the Mars orbit provided much more details. Orbital cameras and infrared spectrometers mapped the spatial and seasonal changes of the dust opacity (Thorpe, 1979, 1981; Martin, 1986). It was found that the storms occur systematically when Mars is near perihelion. Even during the clearest parts of the year, the dust optical depth remains relatively high: $\tau = 0.2–0.4$ in visible wavelengths. Absorption of solar radiation by dust is comparable with that of CO$_2$ gas, which the atmosphere of Mars mainly consists of. Thus, while heating/cooling by CO$_2$ maintains the martian climate, atmospheric dust determines a variability of the circulation and weather, very much like the water on Earth.

We begin with the overview of main features of Mars and martian atmosphere in Section 2, and then consider dust properties and observational constrains that control the dust-induced diabatic heating and cooling of the air in Section 3. An approach to analyzing the global-scale circulation is outlined in Section 4. The influence of dust on the zonal-mean circulation and zonally-asymmetric eddies are discussed in Sections 5 and 6, correspondingly.

## 2. Basic facts about the martian atmosphere

Mars is the second best-studied planet after the Earth. It is often called a “terrestrial-like” planet. Comparison of main physical parameters for Mars and Earth are given in Table 1. Martian radius is about a half, and gravity acceleration is $\sim 0.38$ that of Earth. However, its surface area is only slightly less than the total area of Earth’s dry land. The martian atmosphere is very thin (the surface pressure is about 6 mbar, which is more than 150 times smaller than on Earth), and very dry (there is virtually no water except in subtroughs in the northern summers and in polar regions). It consists mainly of CO$_2$ (95.3%) with other important constituents being N$_2$ (2.7%), Ar (1.6%), O$_2$ (0.15%) and H$_2$O (0.03%). Since the rotation period and orbit inclination of Mars and Earth are similar, the diurnal and seasonal variations are similar as well. Because martian atmosphere is less dense, the difference of temperature between day and night is significantly larger. For instance, the atmospheric temperature near the surface varies between approximately 200 and 260 K at the summer subtroughs, according to the Mars Pathfinder observations (Schofield et al., 1997). Due to the colder temperatures, the atmospheric component of CO$_2$ condenses at the polar regions during winters (e.g., James et al., 1992), and the averaged surface pressure on Mars varies annually within the range of 25%. Although the atmospheric air is thin, winds on Mars are strong. They can lift in the air millions of tons of dust particles during planetary-scale dust storms, which make the atmosphere opaque.

Marsian dust storms have been observed since the end of the 19th century. A comprehensive list of such activities from 1873 to 1990 is presented in Table 1 of Martin and Zurek (1993). Most of the global-scale dust storms occurred near or before southern summer solstice, which is close to the perihelion. They started as regional storms at elevated plateaus located in low- and mid-latitudes, and then expanded in the east-west direction to encircle the planet in 10–20 days, and to last for 50–100 days before decaying. Many regional dust storms have been observed at the edge of receding south polar cap (Briggs et al., 1979), and in northern mid-latitudes during equinoxes (James et al., 1999), all of them being associated with frontal systems. Various mechanisms of dust expansion processes, such as “dust hurricanes” (Gierasch and Goody, 1973), diurnal Kelvin tidal mode (Zurek and Leovy, 1981), have been discussed. In the recent decade, the processes of dust lifting and transport have been studied extensively using martian general circulation models (GCMs) (Newman et al., 2002a, 2002b; Wang et al., 2003, 2005; Basu et al., 2004, 2006; Kahre et al., 2006, 2008).

## 3. Properties of the airborne dust

### 3.1. Seasonal and spatial distributions

Martian atmospheric dust consists of mineral particles eroded from rocks. It is continuously supplied from the martian surface owing to two physical processes: the near-surface wind stress (e.g., Greeley and Iversen, 1985), and small-scale convective vortices (“dust devils”) clearly observed from the Mars Exploration Rover Spirit (Greeley et al., 2006). The amount of airborne aerosol is highly variable with most of the dust storm activity occurring in southern springs and summers. Because of the large orbit eccentricity (0.0934), the planet receives 1.3 times more of solar radiation near the perihelion, which almost coincides with the southern summer solstice (aerocentric longitude $L_s = 270^\circ$). Therefore, the lower atmospheric convection and meteorological processes are more violent during this season compared to northern springs and summers. The scales of dust storms that occur in this period vary from year to year, and range from regional to planet-wide. In the southern spring and summer, the observed maxima of the mean dust opacity over the equator are between 0.25 and 0.5.
1.4 in infrared (9 μm) wavelength (Liu et al., 2003), depending on the scales and strength of the storms. In contrast, a relatively low dust opacity and temperature persist in northern springs and summers, and exhibit high year-to-year repeatability. The measurements from Mariner 9 Infrared Interferometer Spectrometer (IRIS), Viking Infrared Thermal Mapper (IRTM), Thermal Emission Spectrometer onboard Mars Global Surveyor (MGS–TES) and Thermal Emission Imaging System onboard Mars Odyssey (ODY–THEMIS) show that zonal-mean 9 μm dust opacities at the equator were around 0.05 throughout the aphelion season at all years of observations (Liu et al., 2003; Smith, 2004, 2009). A multi-year record of zonally-averaged column dust opacity is shown in Fig. 1. It presents a compilation of data obtained with two instruments: MGS–TES (Smith, 2006) and ODY–THEMIS (Smith, 2009). The record illustrates a wide interannual variability of the dust load during the perihelion season. Dust events differ by the timing of their onsets, duration, strength, latitudinal extent. Particular strong dust storms are seen in MY25 and MY28. Note also a local load during the perihelion season. Dust events differ by the timing of their onsets, duration, strength, latitudinal extent. Particular

3.2. Vertical distributions

The first vertical profile of atmospheric dust on Mars based on Mariner 9 spacecraft limb observations has been reported by Leovy et al. (1972). Seasonal and latitudinal changes of cut-off heights of aerosol distributions have also been obtained from Mariner 9 (Anderson and Leovy, 1978). Jaquin et al. (1986) and Korablev et al. (1993) presented the vertical profiles measured from Viking Orbiter and Phobos spacecraft, respectively. These distributions have been approximated by an analytical formula suggested by Conrath (1975), who assumed that the vertical mixing of particles is determined by the effective diffusivity and gravitational settling:

$$Q_d(z) = Q_{od} \exp \left\{ \nu \left[ 1 - \exp \left( \frac{z}{H} \right) \right] \right\},$$

where $Q_d(z)$ is the dust mass mixing ratio as a function of height $z$, $Q_{od}$ is the value of $Q_d$ at $z = 0$, $H$ is the scale height, and $\nu$ is the ratio of the characteristic diffusion and gravitational settling times. The latter controls the dust cut-off altitude. This formula in pressure coordinates, $p$, has the form

$$Q_d(z) = Q_{od} \exp \left\{ \nu \left[ 1 - \left( \frac{p_0}{p} \right) \right] \right\}, \quad p < p_0,$$

$$Q_d(z) = Q_{od}, \quad p \geq p_0,$$

with $p_0 = 6$ mbar being the global mean surface pressure. Both expressions have been extensively used by GCM modelers for describing dust vertical profiles and calculating the corresponding heating rates (e.g., Pollack et al., 1990; Wilson and Hamilton, 1996; Moudden and McConnell, 2005). The preset values of $\nu$ varied in their simulations from 0.007 in Forget et al. (1999) to 0.01 in Wilson and Hamilton (1996) and Moudden and McConnell (2005) to 0.03 in Pollack et al. (1990). Observations, however, indicate that dust mixing ratios change with height in a more complex manner. In particular, aerosols always extend higher in the equatorial zone, and decay more abruptly towards the poles in both hemispheres (Anderson and Leovy, 1978; Jaquin et al., 1986). To account for this behavior, Forget et al. (1999) adopted a modified version of (3), which was then utilized in many other GCMs (Lewis and Read, 2003; Kuroda et al., 2005; Hartogh et al., 2005):

$$Q_d(z) = Q_{od} \exp \left\{ \nu \left[ 1 - \left( \frac{p_0}{p} \right)^{\frac{z_{max}}{L_s}} \right] \right\}, \quad p < p_0,$$

where the altitude (in km) of the top of the dust layer, $z_{max}$, varies with latitude, $\phi$, and season, $L_s$ as

$$z_{max}(\phi, L_s) = 60 + 18 \sin(L_s - 160°) - 22 \sin^2 \phi$$

Recently, a new wealth of data on vertical dust distribution has been obtained from Mars Climate Sounder onboard Mars Reconnaissance Orbiter (MRO–MCS) (Kleinböhl et al., 2009). It has been observing the martian atmosphere since 2006 by performing limb measurements with global coverage from the surface up to 80 km. Its data significantly complement the limb sounding data of MGS–TES (McConnochie and Smith, 2008; McConnochie et al., 2009). Fig. 2 presents zonally averaged height–latitude cross-sections of dust for different seasons from McCleese et al. (2010). It demonstrates that the vertical distributions are more complex, and differ from those previously observed. The mixing ratios have peaks at 15–25 km over the tropics during much of northern springs and summers, a so-called “high altitude tropical dust maximum” (HATDM) (Heavens et al., 2011a,b). This HATDM structure neither is consistent with the parameterizations used to drive GCM simulations, nor has been reproduced by models which include interactive lifting and transport of dust (e.g., Richardson and Wilson, 2002; Kahre et al., 2006). The mechanism of the ATDM formation in non-dusty seasons is not yet understood. Heavens et al. (2011b) discussed possible roles of topography, convective effects associated with dust devils penetrating across the boundary layer,
and scavenging of dust by water ice. To date, recognizing the origin of the HATDM, and its possible dynamical importance remain challenging topics for investigations of the dust cycle.

3.3. Particle size distributions

The size of dust particles is a key optical parameter that determines absorption and scattering of solar radiation, infrared emission, and, therefore, atmospheric heating/cooling rates. Measurements of dust particle sizes have been performed with Mariner 9 IRIS (Toon et al., 1977), Viking Lander cameras (Pollack et al., 1977, 1979, 1995), Phobos spacecraft (Drossart et al., 1991; Korabiev et al., 1993), and the Imager for Mars Pathfinder (IMP) (Tomasko et al., 1999; Markiewics et al., 1999). These distributions have been analytically approximated either by the modified gamma or log-normal distributions. An example of the former type has the form (Hansen and Travis, 1974):

$$n(r) = \frac{Cr^{-\alpha}}{r_{\text{eff}}^{\alpha-1} \exp \left( -\frac{r}{r_{\text{eff}} \nu_{\text{eff}}} \right)}$$

where $n(r)$ is the distribution as a function of the particle radius, $r$; $r_{\text{eff}}$ and $\nu_{\text{eff}}$ are the effective radius and effective variance, respectively, and $C$ is the normalization constant for adjusting the total size distribution to the number densities. Tomasko et al. (1999) determined $r_{\text{eff}}$ as $1.6 \pm 0.15 \mu m$ and $\nu_{\text{eff}}$ as $0.2–0.5$, which are consistent with the earlier measurements.

Wolff and Clancy (2003) and Clancy et al. (2003) reported the spatial and time variations of $r_{\text{eff}}$ estimated from the MGS–TES spectral data set. They showed that the values of $r_{\text{eff}}$ were between 1.3 and 3 \( \mu m \) at different latitudes and longitudes near the peak of the global dust storm in 2001 ($L_s = 213^\circ$ of the Martian Year 25, MY25), and were smaller ($\sim 1 \mu m$) in the northern spring and summer. The effective particle radius tends to be larger when the dust opacity is high, and the ratio of opacities in solar and infrared wavelengths (a “visible-to-infrared ratio of opacities”) decreases when $r_{\text{eff}}$ grows (Clancy et al., 2003, 2010; Kahre et al., 2008; Elteto and Toon, 2010). This behavior can be explained from the Mie theory using the Viking Lander and Viking Orbiter data (Toigo and Richardson, 2000), and with GCM simulations employing interactive dust transport for particles with several sizes. Larger-size particles produce stronger forward scattering, or weaker extinction by scattering, and this tendency increases for radiation with shorter wavelengths. Thus, the ratio of opacities, an important parameter that affects the radiative heating/cooling, is a function of particle sizes and their distributions.

3.4. Refractive indices

Dust refractive indices are another set of optical parameters that are important for radiative calculations. Presently, most martian GCMs adopt the set of indices obtained by Ockert-Bell et al. (1997) (for wavelength shorter than 5 \( \mu m \)), Toon et al. (1977) (between 5 and 17 \( \mu m \)), and Forget (1998) (longer than 17 \( \mu m \)). Ockert-Bell et al. (1997) inferred the refractive properties at four visible wavelengths (0.5–0.86 \( \mu m \)) from the particle size distribution, shape and single-scattering properties reported by Pollack et al. (1995). They also extended the coverage to all solar wavelengths (0.2–4.2 \( \mu m \)) using ground-based telescope data (Owen and Sagan, 1972; Bell et al., 1990; Rousch et al., 1992), and Phobos-2 ISM data (Mustard et al., 1993) at 0.77–2.9 \( \mu m \). At infrared wavelengths, Toon et al. (1977) compared the Mariner 9 IRIS spectra at 5–50 \( \mu m \) with the terrestrial mineralogical samples and found that montmorillonite 219b, a clay mineral sample with at least 60% of SiO$_2$, has the best fit for the wavelengths shorter than 15 \( \mu m \). For longer waves, Forget (1998) introduced a “synthetic” model to fit the Mariner 9 IRIS spectra.

Wolff and Clancy (2003) suggested a further update for wavelengths between 0.2 and 135 \( \mu m \) based on the set described above and on iterative adjustments using a variety of the MGS–TES measurements. Their distribution was mainly based on a Hawaiian halite sample (Clancy et al., 1995), including a power-law extrapolation in the wavelengths longer than 35 \( \mu m \), and on fitting to the estimates from the IMP observations (Tomasko et al., 1999). Wolff et al. (2006) updated the refractive indices in infrared wavelengths using the upward-viewing geometry of the Miniature Thermal Emission Spectrometer (Mini-TES) on Mars Exploration Rovers. Recently, Wolff et al. (2009) updated the refractive indices in visible to near-infrared wavelengths using the spectra observed by the Compact Reconnaissance Imaging Spectrometer (CIRS) on board Mars Reconnaissance Orbiter during a planet-encircling dust storm in 2007. Fig. 3 shows the refractive indices from (Ockert-Bell et al., 1997; Toon et al., 1977; Forget, 1998) (denoted as “Refractive A”), and from (Wolff and Clancy, 2003; Wolff et al., 2006, 2009). Imaginary parts of the refractive index represent the

Fig. 2. Log$_{10}$ of the zonal-mean dust density-scaled opacity (m$^2$ kg$^{-1}$) observed by the MRO–MCS. After McCleese et al. (2010).
effects of absorption. It is seen that they are larger for the “Refractive A” set in ultraviolet and visible wavelengths (0.2–0.5 μm), and, thus, produce higher daytime temperatures when used in radiative calculations.

4. Approach for analyzing the global circulation

Having discussed the main properties of the dust suspended in the air, we now turn to the effects it produces in the atmosphere of Mars. A convenient way of analyzing dynamics of planetary atmospheres is the concept of wave–mean flow interactions. According to it, all field variables (temperature, pressure, density, wind) are subdivided into the “mean” (spatial, temporal, or combination of both), and deviations. For fast rotating planets, like Earth or Mars, an obvious choice is the separation of the variables into symmetric with respect to the rotational axis (or zonal-mean) ones, and non-zonal disturbances (or eddies). Global-scale distribution of axisymmetric fields is often called the “general circulation”. Due to the non-linearity of the Navier–Stokes equations governing the dynamics, the general circulation is then driven not only by zonally averaged diabatic heating and cooling, but by covariances of eddy fields as well. The latter is called an “eddy” or “mechanical” forcing of the circulation.

In convectively stable atmospheres, many disturbances behave like waves. In other words, they (1) are oscillatory motions that possess a high degree of coherence, (2) propagate horizontally and/or vertically with time, (3) transport energy and momentum, and (4) deposit them to the mean flow upon dissipation. The most prominent large-scale atmospheric waves on Mars are thermal tides and planetary waves.

Solar thermal tides are oscillations of winds, temperature, density, and pressure induced by diurnally varying absorption of solar radiation in the atmosphere and/or surface. They are particularly strong on Mars due to lower than on Earth density, and larger heating rate per unit mass. The solar forcing of tides increases when the amount of dust in the air rises. Westward propagating Sun-synchronous components carrying westward momentum are the direct dynamical response to this forcing. Due to the non-linearity of atmospheric motions, tidal harmonics with various zonal wavenumbers, \( s \), are generated. Vertical and meridional (South–North) propagation of the tides is very sensitive to the mean wind. The diurnal component with \( s = 1 \) can propagate vertically in tropics and poleward in mid-latitudes aloft, but tends to be trapped at low levels. The semi-diurnal Sun-synchronous component (\( s = 2 \)) has longer vertical wavelength, little vertical phase propagation, and depends less on seasons. Variations of topography and surface thermal properties can interact non-linearly with the solar forcing to produce non-Sun-synchronous (or “non-migrating”) tidal modes. Particularly strong on Mars are eastward propagating diurnal (\( s = 1 \)) Kelvin waves, which arise from interactions of the diurnal tide with \( s = 2 \) topography variations (Wilson, 2000; Forbes and Hagan, 2000).

Flow over topography with non-uniform surface thermal inertia generates stationary planetary waves (SPW). These disturbances with observed phase velocities close to zero (but with non-zero speed with respect to the mean flow) can penetrate high into the middle atmosphere if propagation conditions, determined mainly by the mean zonal wind, are favorable. As on Earth, larger-scale martian SPWs with zonal wavenumbers \( s = 1 \) and 2 propagate vertically, whereas smaller-scale ones are mostly trapped near the surface (Banfield et al., 2003; Hinson et al., 2003).

Traveling planetary waves (PWs) are generated by baroclinic and/or barotropic instabilities of the mean zonal flow. They have a broad nomenclature with variety of periods longer than one sol, horizontal wavelengths, and non-zero phase velocities. The spectrum of traveling PWs consists of various well-defined harmonics with certain dominating modes at each time and the transition between them (Hinson, 2006). In the lower atmosphere, they are associated with baroclinic eddies or weather patterns, which can give rise to flushing dust storms (Wang et al., 2003; Basu et al., 2006; Hinson and Wang, 2010). Under appropriate conditions, traveling PWs can propagate into the upper atmosphere (Seth and Rao, 2008), where they contribute to the mechanical forcing of the mean circulation.

5. Influence of dust on the zonal-mean circulation

5.1. Dust storm effects on temperature

In their pioneering work, Gierasch and Goody (1972) first indicated that the observed vertical temperature profiles could not be reproduced in a radiative–convective model without taking into account the absorption of solar radiation by dust. The theoretical studies that followed (Moriyama, 1974, 1975; Zurek, 1978) have demonstrated that the radiative effects of dust are important even when the dust layer is optically thin. The extent to which dust storms modify the atmospheric temperature is illustrated in Fig. 4. It compares the height–latitude cross-sections of the zonal-mean daytime temperature measured by MGS–TES and averaged between \( L_s = 205^\circ \) and \( 210^\circ \) during Martian Years 24 and 25 (MY24 and MY25). A global dust storm occurred in MY25 at this time (see Fig. 1). As a result, the temperature has risen by 40 K at pressure levels 0.2–0.5 mb (15–25 km) over the equator, and more than 60 K over the Sun-lit South pole. This warming has been traced up to 0.01 mb (~60 km) (Gurwell et al., 2005). Conversely, the amplified dust content blocks solar radiation, and

Fig. 3. (a) Real and (b) imaginary parts of refractive indices of martian dust as functions of the wavelength. The values of Wolff et al. (2009) are for \( \tau_{eff} = 1.6 \) μm.
the near-surface temperature drops by \(-10\) K. Such temperature changes alter the convective, baroclinic and barotropic stabilities of the atmosphere, transform wind shears related to meridional temperature gradients, and, thus, produce a dynamical feedback.

5.2. Global meridional circulation

A typical solstitial distribution of zonal mean temperature simulated with a GCM is plotted in Fig. 5. Since the subsolar point, and, thus, the maximum of solar heating is shifted to the southern (summer) hemisphere, the temperature decreases almost monotonically toward the north. In an absence of eddies, such differential heating allows for a development of a thermally direct meridional cell known as a “solstitial Hadley circulation” (Schneider, 1983). It includes rising and sinking air over the warmer and colder hemispheres, correspondingly, a poleward flow due to the conservation of mass, and a return flow near the surface (trade winds). This kind of transport cell, although confined to tropics, is observed in the terrestrial troposphere. The reason for this confinement is that there is no global summer-to-winter-hemisphere temperature gradient because of the very high heat capacity of the oceans. The axisymmetric Hadley circulation is based on the conservation of absolute angular momentum. This implies a negligible role of non-zonal disturbances and their effects, and, therefore, the circulation is sometimes called “almost inviscid”. Being purely thermally driven, the Hadley cell is very sensitive to details of heating. It has been argued (Wilson, 1997; Forget et al., 1999) that on Mars, the circulation of this kind may extend higher and across the globe due to the large pole-to-pole thermal contrast.

The mass-weighted stream function presented in Fig. 5a and b (taken from GCM simulations of Kuroda (2006)) illustrates the changes in the meridional circulation caused by an increased amount of airborne dust. During a relatively dustless season of MY24 (TES2 scenario with \(\tau \approx 0.2\)), the direct thermal cell extends from 30°S to 30°N in the lower atmosphere, and from pole to pole above \(\approx 20\) km. When the dust load dramatically increases to \(\tau = 4.2\) in visible (VIK1 scenario, MY12), the meridional circulation intensifies and broadens. The transport cell expands higher into the atmosphere and extends farther to the winter pole (Fig. 5b). The corresponding zonal-mean temperature distributions are shown in Fig. 5c and d with color shades. It is seen that more dust in the air produces more heating aloft. Consequently, less heating and lower temperature occur near the surface, where solar radiation is weakened being absorbed by the dust. The meridional temperature gradient increases in the Sun-lit summer hemisphere, thus maintaining stronger Hadley circulation. In turn, the axisymmetric meridional winds induce zonal mean westerlies in the winter hemisphere via the Coriolis force, and easterlies in the summer hemisphere and over the tropics (Fig. 5c and d, contours).

Wilson (1997) compared the zonally-averaged meridional mass transport stream function (similar to the one in Fig. 5) with surfaces of the absolute angular momentum. He found that they are roughly parallel over a large region in the atmosphere below \(-0.1\) mb (or \(-40\) km). This implies that the meridional circulation is consistent with the almost inviscid Hadley circulation proposed by Schneider (1983) at these altitudes. In contrast, higher and closer to the poles, significant deviations from the conservation of angular momentum have been found. Together with the well documented departure of the middle atmosphere from the radiative equilibrium, especially over the winter pole, this means that a dynamical forcing by zonally-asymmetric disturbances (eddies) must be invoked.

The effect of eddies on the mean circulation enters the mean momentum equation in the Transformed Eulerian Mean formulation through the divergence of the Eliassen–Palm (EP) fluxes, \(\mathbf{F}\), (Andrews et al., 1987) in the right-hand part:

\[
\bar{u} + \bar{v}'[(a\cos \phi)^{-1}(\bar{u}\cos \phi)_{\phi} - f] + \bar{w}'\bar{u} = (\rho_{0}\bar{a}\cos \phi)^{-1}\nabla \cdot \mathbf{F}. \tag{7}
\]

In (7) and everywhere in this paper, overbars and primes denote longitudinal averages and deviations from the mean, correspondingly; \((\bar{u}, \bar{v}, \bar{w})\) are the components of the zonal-mean wind in \((x, y, z)\) direction, respectively; \(v'\) and \(w'\) are the residual meridional and vertical velocities defined as \(v' = v - \rho_{0}\bar{u}\frac{\partial \bar{u}}{\partial h}\) and \(w' = w + (a\cos \phi)^{-1}(\cos \phi \frac{\partial \bar{u}}{\partial h})_{\phi}\). It is seen, that the net transport represented by the residual velocities is the sum of the Eulerian mean \((\bar{v}, \bar{w})\) and the Stokes drift induced by eddies. In the rest of the notations: \(\phi\) is the latitude, \(\bar{u} = \text{Exp}(Rz/c_{p})\) is the Coriolis parameter, where \(\Omega\) is angular velocity, \(2\pi\) divided by the martian rotational period. It is seen that the divergence of the EP flux (or wave action flux) \(\nabla \cdot \mathbf{F}\) represents a forcing for the zonally averaged components of the wind. \(F\) arises entirely due to the non-linearity of the Navier–Stokes equation when the dynamical variables are split into mean and deviations. In particular, the EP flux depends on covariances between wind and temperature fluctuations and their gradients. It follows from (7) that the eddies not only contribute to the acceleration or deceleration of \(\bar{u}\), but force the meridional transport in the time-average sense (when \(\partial/\partial t \to 0\)).

Using GCM simulations, Wilson (1997) pointed out to a very strong EP flux divergence in the middle atmosphere where the meridional stream functions deviate from absolute momentum surfaces, especially, over the winter pole and during dust storms. Hartogh et al. (2005) and Medvedev and Hartogh (2007) showed that the residual meridional velocity \(v'\) is determined largely by the EP flux divergence at all seasons between 10 and 70 km, at least outside the tropics (\(-30\)° to \(+30\)°). This suggests that the...
meridional circulation on Mars, at least in extratropics, is forced primarily by eddies (planetary waves, tides, and, perhaps, smaller-scale gravity waves) similarly to the so-called “extratropical pump” mechanism (Holton et al., 1995), as on Earth.

5.3. Winter polar warming

A remarkable manifestation of the eddy-driven meridional transport in the middle atmosphere of Mars are temperature inversions and related warmings over the winter poles. They are persistent features in both hemispheres, which usually occur between 50 and 80 km, and have magnitudes from several to tens Kelvin degrees (Deming et al., 1986; Théodore et al., 1993; Santee and Crisp, 1993). Earlier observations have pointed out to a coincidence between the onset of dust storms and polar warmings (Jakosky and Martin, 1987). Since there is no or little of insolation in high winter latitudes, diabatic heating alone cannot cause this effect, even if the dust load is high. It was recognized (e.g., Barnes and Haberle, 1996; Wilson, 1997, and references therein) that polar warmings are associated with the global transport cell, and produced by the descending air entering denser atmosphere. Strong downward motions require strong poleward flow aloft. Thus, the temperature in the middle atmosphere above the polar cap can help to estimate the intensity of the meridional transport, but until recently, most such measurements were restricted to lower altitudes.

Many martian GCMs with sufficiently high model tops reproduced these temperature reversals in high winter latitudes, however, apparently, underestimated their magnitudes. Wilson (1997) was the first to demonstrate that an increase of diabatic heating during dust storms enhances the circulation, and leads to significant warmings. Further modeling studies corroborated this result (Forget et al., 1999; Moudden and McConnell, 2005). Medvedev and Hartogh (2007) explored the dependence between resolved non-zonal disturbances in the GCM, and the simulated residual meridional velocity and winter polar temperature. They argued that the maxima can be ~10–30 K warmer compared to the values predicted by other models (e.g., Forget et al., 1999; Lewis et al., 1999), if simulated tides and planetary waves are consistent with observations and not artificially inhibited in GCMs. Recently, temperature retrievals extending to 80 km have been performed using the MRO–MCS instrument (McCleese et al., 2008). They revealed intense polar temperature inversions in the middle atmosphere (Fig. 6) with magnitudes, which are in a good agreement with the model predictions of Hartogh et al. (2005) and Medvedev and Hartogh (2007). This implies more vigorous than previously expected meridional transport, and, therefore, stronger eddy forcing.

What kind of atmospheric waves contribute most to the EP flux divergences, and why they enhance during global dust storms? To date, the MCS measurements were made over the seasons with relatively low atmospheric dust amount (the data for the 2007 major dust storm were unfortunately missing except for very small regions near the North pole), and, therefore, cannot constrain and/or validate GCM simulations for dust storms. Barnes and

![Fig. 5.](a) Mass stream functions simulated for the “low dust” ($s=0.2$) and “dust storm” ($s=4.2$) scenarios, respectively. Air parcels move along the isopleth so that smaller values remain on the right. (b) The corresponding temperature (shaded) and mean zonal wind (contours). After Kuroda (2006).}
Small-scale gravity waves (GWs) can penetrate high into the middle atmosphere and impose a significant wave drag on the zonal flow upon their breaking and/or saturation. GWs alone are responsible for the temperature inversion and the zonal wind reversal in the terrestrial mesosphere. Effects of orographically generated GWs in the martian atmosphere were studied by Barnes (1990), Joshi et al. (1995), Collins et al. (1997). In simulations of Wilson (1997), the EP flux divergence was dominated by the contribution of thermal tides. With the increase of solar radiation absorption during dust storms, the diurnal variations of temperature are stronger in the atmosphere and surface, and tidal amplitudes are larger (Wilson and Hamilton, 1996). Kuroda et al. (2009) analyzed the eddy forcing of the vigorous poleward and downward transport during a planet-encircling dust storm in detail. They showed that the warming is caused by dissipating thermal tides, planetary and resolved-scale GWs in almost equal degree. In their simulations, one third of the EP flux divergence was contributed by solar tides excited in the summer hemisphere, especially around the subsolar latitude. The other two thirds of the required forcing were tides excited in the summer hemisphere, especially around the subsolar latitude. The other two thirds of the required forcing were created by the stationary planetary wave (SPW) with the zonal wavenumber $s=1$, the transient planetary wave ($s=1$, period ∼5 sol), and the resolved GWs.

A similar polar warming but in the lower thermosphere at 100–130 km have been found during the ODY aerobraking phase around $L_s = 270^\circ$ (Keating et al., 2003). GCM simulations of Bougher et al. (2006) demonstrated that the mechanism of the warming formation is analogous to the one in the middle atmosphere. Using a lower-to-middle atmosphere GCM coupled with a thermosphere model, Bell et al. 2007 and McDunn et al. (2010) have shown that the meridional transport in the thermosphere intensifies, and the polar temperature grows when the total amount of dust increases and/or it is mixed to higher altitudes. Examples of lofting of lower atmospheric aerosols to altitudes as high as 80 km during 2001 strong dust storm event have been observed with the MGS–TES (Clancy et al., 2010). In their simulations with the first “ground-to-exosphere” martian GCM, González-Galindo et al. (2009) emphasized the contribution of both migrating and non-migrating tidal components to the EP flux divergence that drives the lower thermosphere polar warming.

### 5.4. Atmosphere superrotation

We discussed above the main features of the solstitial circulation in the atmosphere of Mars. Over equinoxes, the subsolar point moves to the tropics giving rise to almost symmetric with respect to the equator temperature distribution in the lower atmosphere. Being in transition between two solstices, the equinoctial zonally averaged circulation consists of two meridional cells with rising motions over the equatorial region and the downward flow over the poles. This structure reproduced by all martian GCMs extends up to at least 70 km or higher. As a result of the Coriolis force, two westerly (prograde) jets are maintained by the meridional cells in mid- and high-latitudes in both hemispheres. In low-latitudes, where the Coriolis force and the cross-equatorial flow are weak, the zonal wind is more influenced by eddies (see Eq. (7)). In the absence of forcing associated with non-axisymmetric disturbances zonal wind remains purely easterly (retrograde) (Hide, 1969). Thus, the superrotation, defined as a ratio of the positive (prograde) angular momentum of the atmosphere and that of the atmosphere at rest, depends on the eddy forcing.

It has been noticed that the atmosphere of Mars can exhibit a local superrotation under dusty conditions (Hamilton, 1982; Zurek, 1986; Wilson and Hamilton, 1996). Lewis and Read (2003) have found that the superrotation is most pronounced in the lower equatorial atmosphere (one half to one scale height above the surface) at equinoxes. They showed that the equatorial westerly jet can extend above ~40 km during strong dust storms. In agreement with the work of Wilson and Hamilton (1996), solar tides have been identified as the main driving force of the low-level westerly jet. When the dust opacity is large and the absorption of solar radiation increases, temperature swings between day and night become particularly strong. They amplify the excitation of solar tides, mainly of sun-synchronous diurnal and semi-diurnal modes. These tides provide a superrotation torque at places of their generation, in accordance with the mechanism proposed by Fels and Lindzen (1974). This can be illustrated with the generalized version of the Eliassen and Palm theorem:

\[
\frac{d}{dz} \left( \rho_0 \overline{\nu \nabla W} \right) = - \frac{K \rho_0}{u - c} \overline{\nabla F},
\]  

(8)
where $c$ is the phase velocity of the disturbance, $\kappa = (\gamma - 1)/\gamma$, $\gamma$ being the ratio of specific heats $c_p/c_v$, $D = (\zeta/H + (p'/p')$ is the measure of a vertical displacement in the wave, $\zeta$ is the vertical displacement of a parcel, and $J$ is the heating rate. The divergence of the eddy momentum flux in the left-hand side of (8) enters the right-hand side of (7), and represents a forcing for the mean zonal wind. It is seen from Eq. (8) that it acts to accelerate the flow in a direction opposite to $c$. Thus, westward-propagating migrating tides produce eastward accelerations in regions where heating takes place, extract westward angular momentum from the mean flow, transport it upward, and deposit it back upon their dissipation. Both the low-level westerly flow and easterlies aloft the equatorial region are directly related to the thermal excitation of tides. Lewis and Read (2003) found that the superrotation of the martian atmosphere increases slowly with dust loading in integral sense, and strongly depends on the latter locally.

5.5. Equatorial semiannual oscillations

Another manifestation of the wave-induced superrotation in the martian atmosphere is the semiannual oscillation (SAO) of the mean zonal wind in tropics. The SAO is well documented in the stratosphere of Earth (Garcia et al., 1997), and has been recently found on Saturn as well (Orton et al., 2008). Kuroda et al. (2008) detected a semiannual periodicity in the difference between the day- (local time $\sim 2:00$ pm) and night-time (local time $\sim 2:00$ am) temperature retrieved from the MGS–TES measurements. The contour interval is 10 m s$^{-1}$, westerly (prograde) wind is shaded with yellow. (b) Same as Fig. 7a, except from the run with the seasonally uniform dust opacity in visible wavelengths $\tau = 0.2$. After Kuroda et al. (2008).

6. Influence of dust on non-zonal disturbances

6.1. Thermal tides

A significant fraction of the natural variability in the Mars middle atmosphere is associated with the thermal tide activity. They are composed of in-situ generated tides, and harmonics propagating from below. Diurnal and semidiurnal Sun-synchronous (“migrating”) tides are the two main modes of the atmospheric response to the daily cyclic absorption of solar radiation. The semidiurnal tide has a much longer vertical wavelength than its diurnal counterpart, and therefore, is less subjected to vertical damping by the background zonal winds. If heating becomes distributed over greater depths in the atmosphere, the efficiency of exciting semidiurnal tides increases. Therefore, during dust storms, the semidiurnal tide has noticeably larger amplitudes, extends well above the region of excitation, and dominates the diurnal component almost everywhere in the middle and upper atmosphere. This behavior has been well reproduced in GCMs (Wilson and Hamilton, 1996; Wilson and Richardson, 2000) and quasi-linear tidal models (Forbes and Miyahara, 2006). However, diurnal temperature perturbations can be quite large too ($\sim 20–45$ K) within the heating region during planet-encircling dust storms (Smith et al., 2002). These variations are composed of vertically trapped diurnal modes with very long vertical scales. They contribute to the enhancement and poleward extension of the meridional circulation, as was discussed above.

Above $\sim 100$ km, both molecular viscosity and convective instability compete for limiting exponentially growing (due to density stratification) amplitudes of vertically propagating tides. Forbes and Miyahara (2006) have estimated that, above 150 km, amplitudes of semidiurnal temperature fluctuations, although being large, do not vary significantly (from 50 to 70 K) when the dust opacity rises from low ($\tau = 0.5$) to very high ($\tau = 5.0$). This occurs due to the dissipation in the thermosphere, and amplitude saturation. Between 50 and 100 km, the amplitudes are more sensitive to the amount of airborne aerosol.

6.2. Traveling planetary waves

Traveling planetary waves with periods from 2 to 10 sol were observed in the northern hemisphere of Mars from autumn to spring. Occurrences of dust storms seriously modify excitation.
and propagation characteristics of traveling PW, and, thus, affect their periods, zonal wavenumbers, and amplitudes. In particular, ejections of large quantities of aerosol to the atmosphere are accompanied by a significant reduction of wave activity. Fig. 8 (red line) shows the time series of pressure variations at the surface (47°N) measured by Viking Lander 2. They are the result of density oscillations created by traveling PW in the atmosphere above. A noticeable decrease of the wave activity (left panel) has been reported in observations (Barnes, 1980) and GCM simulations (Kuroda et al., 2007) after the onset of a planet-encircling dust storm with global mean visible dust opacity up to ~4.2 at \( L_s \sim 280^\circ \), which lasted for 30–40 sol. These changes are caused by the baroclinic stabilization of the flow. Fig. 5d shows that the low-level vertical wind shear in mid-latitudes of the northern (winter) hemisphere decreases as a result of the meridional temperature gradient flattening. The weaker wind shear stabilizes the flow with respect to baroclinic perturbations, especially for longer-scale harmonics. This mechanism serves as a negative dynamical feedback for dust storm developments. The enhanced mean meridional circulation and reduced baroclinic wave generation cause a decline of dust particle lifting from the surface (Basu et al., 2006; Kahre et al., 2006).

7. Summary and conclusions

Diabatic heating due to absorption of solar radiation by airborne dust particles, and the associated cooling due to emissions in infrared are the prominent factors driving the atmosphere of Mars at various scales. Our paper provides a survey of relatively recent investigations on the influence of dust on global circulation and large-scale motions. We focus on properties of the dust which determine the diabatic heating/cooling, and on dynamical feedbacks to it.

There is a variety of meteorological phenomena in the lower atmosphere which involve dust. Following the term introduced by Heavens et al. (2011b), vigorous conimeteorological (or “dust meteorological”) processes take place within the lower scale height (~10 km), which lift the dust from the surface. As a result, suspended particles may remain in the martian atmosphere for months. Observational data from the spacecraft in the last decade and broad modeling efforts provide an evidence that, although dust heating/cooling is local, its effect is global.

We outlined in this review the main dust-related dynamical effects including the influence on the zonal-mean circulation, and on non-zonal disturbances: planetary waves and tides. There are many open questions in this research area that still remain. Some immediate of them are:

- How high the response to dust storms is transmitted in the atmosphere? There are indications that strong dust events in the lower atmosphere may affect the ionosphere (Haider et al., 2010). What are the coupling mechanisms? Are they purely dynamical or include interactions with chemical and electrodynamical processes?
- Non-migrating tidal signatures are persistent in the upper atmosphere (Forbes, 2004), and have large amplitudes. The associated density fluctuations in the thermosphere are strong enough to affect aerobraking of spacecraft. How does atmospheric dust influence non-migrating tides?
- All martian GCMs consider that dust consists of particles with one or few sizes only. As observations show, this is not true. What are the dynamical consequences of accounting for the variable dust size?
- We did not consider in this review the dynamical influence of gravity waves, and its relations with a variable dust load. The only reason for this omission is that this topic is still underdeveloped. However, there are indications that GWs are extremely important in the thermosphere (Medvedev et al., 2011). Since propagation of GWs is very sensitive to the mean zonal wind, what is their response to the dust-modified background?

Although this review does not touch upon the role of aeolian processes near the martian surface, the pertaining research is actively advancing (Basu et al., 2006; Kahre et al., 2006, 2008; Newman et al., 2002a; Wang et al., 2005). The most challenging goal is to understand why some of localized dust lifting events develop into local-area dust storms, and how they strengthen to large-scale and planet-encircling events. This also includes an explanation for the quasi-regularity of dust storms and their initiation in certain areas. The mentioned above fundamental science question is becoming a practical as well: an accurate weather prediction is required for planning numerous martian missions that include landers and rovers. Most likely, a combination of geological, aeolian, meteorological and circulation processes are responsible for the dust storm phenomena. This is a fast growing field of martian atmosphere research where the aeolian science community can greatly contribute.

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