Simulation of the 2018 Global Dust Storm on Mars Using the NASA Ames Mars GCM: A Multitracer Approach

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Abstract Global dust storms are the most thermodynamically significant dust events on Mars. The most recent of these events occurred in 2018. Although it was monitored by several spacecraft in orbit and on the surface, many questions remain regarding its onset, expansion and decay. Here, we model the 2018 event with the National Aeronautics and Space Administration (NASA) Ames Mars Global Climate Model in order to better understand the evolution of the storm. Our results highlight a mechanism for the expansion of the storm: the initial equatorial regional storm creates a zonal atmospheric temperature gradient causing strong equatorial eastward winds and thus rapid eastward transport of dust and subsequent lifting. The model shows rapid back and forth transfer of dust between western and eastern hemispheres reservoirs, which may also play an important role in the storm's development through teleconnections involving replenishment of surface dust. The model also shows that gigantic dust plumes occur during the storm's mature phase, injecting dust up to 80 km. Our analysis shows that their upward motion in the atmosphere is due to the ascending branches of Hadley cells, whose intensity is reinforced during the storm with increasing dustiness. We show that the global atmospheric warming during the storm cause vapor and water ice clouds to migrate to higher altitudes, in line with recent observations. Finally, we find that the choice of effective radius for the lifted dust particle size distribution impacts the intensity of the Hadley circulation and could explain some of the differences obtained between model results and observations.

Plain Language Summary Global dust storms are planetary-scale events on Mars but remain difficult to predict with climate models because the mechanisms of their onset and evolution are not well known and involve many subtle positive or negative feedbacks between the circulation and heating. The most recent of these events, which began in June 2018, was monitored by several spacecraft in orbit and on the surface. Here, we model this global dust storm with the NASA Ames Mars Global Climate Model to better understand the evolution of the storm. We find that the global dust storm is characterized by a rapid eastward transport of dust in the equatorial regions and subsequent lifting. We highlight rapid back and forth transfer of dust between western and eastern hemispheres reservoirs, which may play an important role in the storm development through long-distance connections between regions involving replenishment of surface dust. We also investigate large dust plumes occurring during the mature phase of the storm and injecting dust up to 80 km. Finally, we find that the height at which water condenses to form clouds increases during the storm, leading to more water vapor in the upper atmosphere.

1. Introduction

Global dust storms (GDSs) are the largest spatial-scale dust lifting events on Mars and represent one of the most puzzling phenomena of the Mars dust cycle. They occur only every few Martian years (MY, see definition in Clancy et al., 2000), during Mars' dusty season (i.e., northern fall and winter, corresponding to solar longitudes \( L_e = 180°–360° \)) and usually mask most of the surface from orbit for several months as they inject large amounts of dust into the atmosphere and produce high dust visible optical depths (typically larger than 3 and up to 10). Global storms appear to have a distinct climatology from the more regular seasonal cycle of regional storms, which includes the three dust activity seasons categorized by Kass et al. (2016): the presolstice A, solstice B, and postsolstice C seasons.

Following the start of global mapping in 1999/MY24 by Mars' Global Surveyor (MGS), Mars has been under near-constant surveillance by one or more spacecraft up to the present (MY35), providing detail on the impact of dust storm activity on the Martian climate. Dust storms and the subsequent atmospheric dust loading significantly warm the atmosphere of Mars and thus alter the atmospheric circulation as well as
Figure 1. The seasonal variation in zonally averaged equatorial (10°S–10°N) $T_{15}$ from 3 pm. MCS limb observations for Mars Years 28 to 34. These MCS temperatures are based on 15-μm brightness temperatures in limb viewing mode, representing a depth-weighted layer of atmosphere centered at ~30 Pa. Also shown are TES temperatures at 30 Pa for Mars Years 24 to 26. The GDSs are significant outliers in the normal cycle of A-season ($L_s = 210–235^\circ$) and C-season ($L_s = 310–340^\circ$) regional storm activity. The B-season storms do not show up on this figure because they remain confined poleward of the southern midlatitudes. Note that other presolstice and postsolstice GDSs have been confirmed during MY 9, 10, and 12 (e.g., Martin; 1974, Zurek & Martin, 1993).

The CO$_2$ and water cycles (e.g., Guzewich et al., 2014; Kahre et al., 2017; Newman et al., 2002; Strausberg et al., 2005; Wilson, 1997). Figure 1 shows the last 10 MY of equatorial zonal-mean $T_{15}$ temperatures, representative of a depth-weighted layer of atmosphere centered at ~30 Pa (see details on $T_{15}$ in section 2.1), and demonstrates the particularly impressive impact of global storms on atmospheric temperatures. Global storms also have a global impact on surface properties. For instance, changes in surface albedo have been observed after the occurrence of GDSs (e.g., Cantor; 2007; Zwast et al., 2006; Vincendon et al., 2015), suggesting a redistribution of the surfact dust reservoirs as dust deposits tend to brighten the surface. These transfers of dust from different surface reservoirs are key as they may create the hysteresis responsible for the inter-annual variability of GDSs (Kahre et al., 2005; Newman & Richardson, 2015).

The most recent GDS, the 2018/MY34 GDS, was observed by several instruments, including the Mars Color Imager (MARCI) on-board Mars Reconnaissance Orbiter (MRO), which produced daily global maps of Mars during this period (Malin et al., 2018a, 2018b, 2018c, 2018d, 2018e, 2018f, 2018g), and by the Mars Climate Sounder (MCS), which produced profiles of temperature and aerosol (Montabone et al., 2020, this issue). This GDS, like that in MY25, is characterized by a relatively early season of onset (shortly after the northern fall equinox, $L_s = 180^\circ$) and amplitudes in opacity and temperature similar to most GDSs. Modeling studies of GDSs with parameterized dust lifting easily produce storms in the solstice season (onset around $L_s = 250–280^\circ$). This is because the Hadley circulation, which is most intense in this season, triggers efficient lifting, transport of dust, and associated positive feedbacks thus driving the genesis of the GDSs in the models (e.g., Basu et al., 2004, 2006; Haberle et al., 1982; Kahre et al., 2005; Newman & Richardson, 2015). However, the 2007/MY28 storm, which began at $L_s = 265^\circ$ (Montabone et al., 2015; Wang & Richardson, 2015), is the only solsticial storm seen to date in the era of continuous spacecraft monitoring (MY23–MY34). By contrast, both the 2001/MY25 and 2018/MY34 events began soon after $L_s = 180^\circ$ (e.g., Cantor, 2007; Guzewich et al., 2019). This points to other factors being important to global storm onset. For example, Newman and Richardson (2015) simulate GDS onset as early as $L_s = 210^\circ$ in their model when surface dust abundances are assumed to be finite, which changes the primary dust lifting source regions and timings. Also, Kahre et al. (2011) simulate GDS onset as early as $L_s = 180^\circ$ when water ice microphysics and radiative effects are included. Yet how such simulations relate to Mars itself—in other words, which process (or combination of processes) is truly most important—is not currently well understood.

Although the 2018/MY34 GDS was monitored by several instruments in orbit and on the surface of Mars, many fundamental questions remain unanswered regarding its onset, evolution, and impact on the Martian climate. In particular, what controls and triggers the onset of the GDS? How does the GDS expand, and why and by which mechanisms does it stop expanding? Where are the surface dust reservoirs and is the transfer of dust between reservoirs important to how the storm evolves? What are the differences between the MY34 and MY25 GDSs events, and how are these two storms distinguished from the regular regional storms that develop in the A-season ($L_s = 210^\circ–235^\circ$) and from the less regular and weaker regional storms that occur around equinox in some years (Battalio & Wang, 2019; Wang & Richardson, 2015)? How rapidly are finite dust reservoirs depleted and restored? What could be the teleconnections between lifting sites (i.e., how a dust event in one region could affect weather pattern and dust lifting in another region at distance)? These scientific questions call for modeling efforts of the MY34 GDS.

In this paper, we aim to obtain new insights on these questions and more generally on the evolution of the present-day dust cycle on Mars by simulating the MY34 GDS with the NASA Ames Mars Global Climate Model (MGCNM). We use observed dust column abundances to drive dust lifting only, with the model then self-consistently simulating dust transport/sedimentation and the general circulation. This allows us to investigate feedbacks between dust lifting, surface stresses, and aerosol transport in the simulation, which
provides valuable insight into what may have happened during the real GDS. Here, our goals are to capture the storm evolution; qualitatively and quantitatively assess the sources, sinks, and transport of dust; study the sensitivity of the circulation to different parameters; and further our understanding of the dust cycle on Mars. In addition, the MY34 event presents an excellent opportunity to assess the MGCM’s new “tagging” method, which allows dust to be tracked using a broad range of selection criteria, including spatial and/or temporal origination, type of lifting process, and so forth (see sections 3.2 and 3.3.3).

We first describe the observational metrics used in this paper (MCS brightness temperatures and MCS‐derived daily global dust maps [DGDM]) and give an overview of the MY34 GDS in section 2. We then present the model used and describe the settings of the reference simulation in section 3. Section 4 gives an overview of the MY34 GDS as simulated by our reference simulation, compares the results with the available observations, and discusses the discrepancies. Section 5 provides an analysis of the dust sources, sinks, and transport obtained in this simulation. Finally, in section 6, we discuss our results further in light of a sensitivity study and comparisons with other regional and GDSs.

2. Background: Observations of the MY34 GDS

In this section, we give a brief overview of how the 2018/MY34 GDS evolved based on the available observations (MARCI and MCS). Other descriptions of the storm can be found in Guzewich et al. (2019), Sanchez‐Lavega et al. (2019), Kass et al. (2019), and in further detail in other papers of this special issue. The locations of the different regions cited in this paper are shown by Figure 2.

2.1. Description of the Set of Observations

2.1.1. DGDM

Spacecraft observations have enabled the creation of a daily record of 3.75° × 3.75° spatially resolved column dust infrared (IR) opacity from mostly the Thermal Emission Spectrometer (TES), on-board MGS, and MCS observations for the last 10 Martian years (Montabone et al., 2015). In the case of the MGS mission, these fields are primarily derived from TES nadir observations of column opacity. For MRO, the column opacity is primarily derived from the downward extrapolation of profiles of dust opacity retrieved from limb viewing MCS observations.

These sets of gridded, daily dust opacity maps are an important input for MGCM simulations (see section 3) and will be referred to in this paper as DGDM. Here, we use the most recent DGDM for MY34,
derived from the available opacity observations provided by MCS. The version of these DGDM is V2-5 (Kleinböhl et al., 2017; Montabone et al., 2020). We have converted the IR (9.3 μm) opacities to visible (0.67 μm) opacities for comparison with visible opacities typically reported by MGCM modeling groups, using a fixed opacity ratio of 2.6 (1.3 times 2 for the ratio VIS to IR and absorption to extinction, respectively, as in Montabone et al., 2020). Note that this factor assumes a fixed particle size distribution, which implies some uncertainties as the particle size distribution is expected to vary with time, in particular during GDSs.

2.1.2. \( T_{15} \) Temperatures

In this paper, in order to illustrate the impact of dust opacity on global scale atmospheric temperature, we compare simulated and observed brightness temperatures. The observations are derived from MCS radiances from the lowest detector in the 15 μm (A3) channel included in the level 1 data product in the Planetary Data System archive. We refer to this measure as \( T_{15} \), which represents a depth-weighted temperature centered at 40 Pa (~30 km) where the A3 channel weighting function peaks. The on-planet vertical weighting function is similar to that of the Viking IRTM 15-μm channel (Wilson & Richardson, 2000). The use of brightness temperature bypasses the temperature retrieval process required for profiling and yields results that are relatively uninfluenced by aerosols (since the dominant opacity contribution at 15 μm is CO₂). Thus this measurement allows relatively low altitude nighttime and daytime temperatures (MCS observations are obtained from a sun-synchronous orbit, providing observations at 3 am and 3 pm mean solar time; Shirley et al., 2019) to be obtained for all latitudes throughout the MRO mission (note that the measure of \( T_{15} \) is not dependent on a successful retrieval of MCS temperature, which sometimes fail in highly dusty circumstances). This is particularly useful for examining the temperature response to dust storm events, as the diurnal temperature range and decay of average temperature is diagnostic of the influence of dust heating (Conrath, 1975). In other words, the measure of \( T_{15} \) is a good indicator for dust heating.

2.2. Overview of the MY34 GDS

The 2018/MY34 GDS lasted about 110 sols. The description of its temporal evolution will vary depending on whether we consider its impact on opacity, temperature, or pressure and whether we consider its impact globally or at a particular location (e.g., a spacecraft landing site). If we consider its thermal impact on equatorial temperatures (shown in Figure 1), however, the GDS started as a cross-equatorial event on 2 June (Lₐ ~ 186°, just after northern spring equinox), became “planet-encircling” around 20–22 June (Lₐ ~ 197°), and achieved peak global opacity in the first half of July (Lₐ ~ 205°–210°), after which there was a gradual decay before typical background dust levels were reached at the end of September (Lₐ ~ 270°). As observed during previous GDS events (MY25 and MY28, e.g., Strausberg et al., 2005; Wang & Richardson, 2015), the MY34 GDS was not composed of a single lifting event but rather was a sequence of several local and regional storms.

MARCI observations revealed several local frontal-like dust storms occurring along the seasonal cap edge in the high-latitude plains of the northern hemisphere around Lₐ ~ 181°. These frontal storms, driven by baroclinic activity (Wang et al., 2003; Wang & Richardson, 2015), are likely to be the precursors of the MY34 GDS. At this season, the general circulation is characterized by a two-cell Hadley circulation with a narrow ascending branch near the equator and broader descending branches at high latitudes. The descending branch in the high-northern latitudes confined the frontal dust storms to the lowest atmospheric levels and transported the lifted dust towards southern latitudes.

Around Lₐ ~ 185°, the frontal storms penetrated the tropics in Acidalia/Chryse Planitia (30°N–60°N, 300–360°E), one of the three low-topography “flushing” channels known to be efficient at transporting dust into the south hemisphere (Utopia/Isidis and Arcadia/Amazonis being the others) and to be an active region for dust lifting and onset of regional storms (Wang & Richardson, 2015). Note that this is the first GDS of the post-1999 “spacecraft monitoring” era that was observed to initiate in the northern hemisphere (MY25 and MY28 initiated in the southern hemisphere, in Hesperia Planum and Chryse/Noachis, respectively). Around Lₐ ~ 187°, the accumulation of atmospheric dust in Acidalia/Chryse Planitia formed a regional storm, which was responsible for the sudden and abrupt increase in the observed global mean dust opacity and atmospheric temperatures then (Figure 1).

From there, the storm expanded eastward and southward. Distinct dust storms also occurred along the receding southern seasonal CO₂ polar ice cap (possibly following a lifting mechanism described in Toigo
et al., 2002), in particular near the region south of Hellas (Ls = 188°–192°), as revealed by MARCI observations (Malin et al., 2018a, 2018b, 2018c, 2018d, 2018e, 2018f, 2018g) and to some extent by the MCS-derived DGDM (Montabone et al., 2020, this issue). By Ls = 192°, following northward expansion, these storms merged with the larger regional tropical storm, which then expanded further east and became global by Ls = 196°. Note that this eastward expansion in the tropics is a dominant feature of the MY25 and MY34 dust storms, which is not the case for the regional storms observed during the dusty season (including the A, B, and C season). From theoretical arguments, the conservation of angular momentum should lead to strong extratropical prograde (eastward) jets and a weak equatorial retrograde (westward) jet in the atmosphere (if one assumes no friction with the surface before the air rises up, which would lead to perfect conservation of angular momentum). However, in the tropics, diurnal thermal tides tend to force an eastward zonal-mean flow at about 10–20 km above the surface (Lewis & Read, 2003; Wilson & Hamilton, 1996). This flow intensifies with stronger dust forcing such as that induced by the onset of the GDS.

As dust lifting continued, dust visible opacities locally peaked between 5 and 10. The global mean dust opacity reached a peak of ~4 around Ls = 205°. Note that this eastward expansion in the tropics is a dominant feature of the MY25 and MY34 dust storms, which is not the case for the regional storms observed during the dusty season (including the A, B, and C season). From theoretical arguments, the conservation of angular momentum should lead to strong extratropical prograde (eastward) jets and a weak equatorial retrograde (westward) jet in the atmosphere (if one assumes no friction with the surface before the air rises up, which would lead to perfect conservation of angular momentum). However, in the tropics, diurnal thermal tides tend to force an eastward zonal-mean flow at about 10–20 km above the surface (Lewis & Read, 2003; Wilson & Hamilton, 1996). This flow intensifies with stronger dust forcing such as that induced by the onset of the GDS.

As dust lifting continued, dust visible opacities locally peaked between 5 and 10. The global mean dust opacity reached a peak of ~4 around Ls = 205°–210°, which corresponds to the maximum thermal impact of the storm, based on tidal analysis at MSL site and observed MCS T13 temperatures and MCS-derived DGDM (e.g., Montabone et al., 2020; Viúdez-Moreiras et al., 2019, this issue). After Ls = 210°, the GDS entered the decay phase and dust started settling out of the atmosphere. Atmospheric dust loading returned to nominal seasonal levels by Ls ~ 250°.

3. Model Description

3.1. The NASA Ames Mars GCM

We use the NASA Ames Mars GCM (MGCM), which now employs (1) the NOAA/GFDLcubed-sphere finite-volume (FV3) dynamical core and (2) physics packages from the Ames Legacy MGCM as described in Kahre et al. (2018) and Haberle et al. (2019). The cubed-sphere grid is relatively uniform, which enables efficient high-resolution simulations on massively parallel computers. The model uses topography from the Mars Orbiter Laser Altimeter and albedo and thermal inertia maps derived from Viking and MGS TES observations. The surface roughness is temporally and spatially fixed to 0.01 m in the simulations discussed in this paper.

The model includes coupled dust and water cycles, using the Ames water ice cloud microphysics package described in Haberle et al. (2019). This includes water sublimation from the north polar residual water ice cap and the complex processes of cloud microphysics (nucleation, growth, and settling) (Montmessin et al., 2002, 2004; Navarro et al., 2014; Nelli et al., 2009). The airborne dust that interacts with solar and infrared radiation acts as ice nuclei and goes through gravitational sedimentation as free dust and as cores of water ice cloud particles. The log-normal particle size distributions of dust and clouds are represented by a spatially and temporally varying mass and number and a constant effective variance (two-moment scheme). Many different dust lifting schemes are implemented, based on observations or equations representing the processes of convective (dust devils) and wind stress lifting (“interactive dust lifting”, e.g., Kahre et al., 2006; Kahrea et al., 2015). However, in this work, we instead impose dust lifting based on the observed column opacity fields, as given by the DGDMs. This is sometimes referred to as “assimilated dust lifting” (e.g., Greybush et al., 2012; Kahre et al., 2009) and is described in detail in section 3.3.2.

The planetary boundary layer (PBL) model in the MGCM solves energy and momentum equations to estimate wind and temperature profiles in the atmosphere and mix tracers, as in Haberle et al. (2019). It employs a Mellor–Yamada level 2 boundary layer scheme for turbulence closure. This implementation is first described in Haberle et al. (1993) and later updated in Haberle et al. (1999). The model calculates surface fluxes based on heat and momentum drag coefficients from the stability functions of Savijarvi (1995) and Hourdin et al. (1995). Eddy mixing coefficients are calculated using the equations of Arya (1988).

The model employs a two-stream radiative transfer scheme that accounts for gaseous absorption of CO2, H2O, and scattering aerosols, including dust and water ice particles (Haberle et al., 2019; Toon et al., 1989). Gaseous opacities are calculated using the correlated-k method, and Rayleigh scattering is calculated for CO2. Mie theory is used to calculate extinction efficiencies and scattering properties for aerosols, assuming a log-normal size distribution. Ice is assumed to have a core-mantle structure, with refractive indices for dust
and ice taken from Wolff et al. (2010) and Warren (1984), respectively. We note that the optical properties of both dust and water ice cloud particles depend on particle size and thus evolve in time and space. In the case of dust, the assumed lifted dust particle size distribution affects key aspects of the radiative behavior of the airborne dust grains (e.g., the visible to infrared ratio). This issue is discussed further in section 4.1. The optical properties are used in a two-stream code to calculate fluxes, and heating rates are computed from the flux divergences. There are seven visible bands (0.4–4.5 μm) and five IR bands (4.5–1,000 μm).

For comparison with MCS $T_{15}$ temperatures, we have synthesized comparable temperatures from MGCM simulation results, in a manner similar to that described in Wilson and Richardson (2000). We use a weighting function appropriate for the off-limb viewing geometry for MCS. We compute radiiances from MGCM columns of temperatures, weight these with the weighting function, and then convert the resulting weighted radiancé back to a brightness temperature.

3.2. The Tagging Method

The MGCM includes a numerical tagging method (Bertrand et al., 2018), which “tags” (or labels or follows) any atmospheric constituent (the “reference tracer”, e.g., dust particles, water vapor, water ice, argon etc.) according to a chosen criterion (e.g., location, local time, type of lifting, amplitude of the dust source or wind stress, altitude attained, etc.). Each tag is transported by the circulation and behaves as the constituent it follows but remains completely passive and does not alter the predictions. This technique enables us to track not only the origin of a given atmospheric constituent but also the physical processes it goes through (e.g., scavenging, formation of ice clouds, frost, etc.), or the different environments it has encountered since its emission (craters, mountains, dusty atmosphere, poles, etc.). This powerful method was first implemented in the NASA/GISS GCM (Koster et al., 1986) to identify the origin of precipitation in various regions of the Earth and is now widely used for detailed studies of the Earth’s water cycle. It has never before been used for Mars.

3.3. Reference Simulation Settings

3.3.1. Overview

The simulations described in this paper are carried out with a horizontal resolution of $2^\circ \times 2^\circ$ and 46 vertical levels, with a vertical resolution decreasing from 20 m near the surface to 10 km at the model top (~80 km). The reference simulation is performed with two-moment aerosol size distributions for water ice and dust. Dust particles are injected from the surface following a log-normal particle size distribution with an effective variance of 0.5 and an effective particle radius of 3 μm (Clancy et al., 2003; Kahrea et al., 2015; Wolff & Clancy, 2003). In the atmosphere, the distribution varies in space and time through the effects of transport and sedimentation. Here, the effective particle radius of the surface injection is larger than the 2.5 μm value previously used in Kahrea et al. (2015). This choice is driven by better agreement between simulated and observed opacities, as highlighted in the sensitivity study detailed in section 6. We assume a dust particle density of 2,500 kg m$^{-3}$.

The simulations are carried out over multiple annual cycles (using DGDM for MY33) so that the aerosol and temperature distributions could reach a seasonally equilibrated state before running the simulation for MY34. We use radiatively active dust, water vapor, and water ice clouds.

3.3.2. Dust Lifting Scheme

There are several possible strategies for dust lifting used for controlling and influencing the 3-D distribution of aerosol in a MGCM simulation. Here, we use the DGDM, interpolated in time and onto the GCM grid, to allow the GCM to identify dust lifting centers and reproduce realistic opacities and temperatures. Dust is injected from the surface (we assume an infinite reservoir across the globe) into the PBL at each physics scheme time step when the simulated dust column opacity is lower than that in the prescribed DGDM (Bertrand et al., 2019; Greybush et al., 2012). The difference in opacity (the observed opacity from the DGDM, interpolated to that time step, minus the GCM simulated value at that time step) is converted to a mass of dust to be injected following eq. (1) in Wang et al. (2018). Typical masses of injected dust during the 2018 GDS are discussed in section 4.4. Dust is then allowed to be transported elsewhere by the simulated general circulation. We allow dust to be lifted at night but not over the polar caps. Note that in this reference simulation, dust is only removed from the atmosphere to the surface by sedimentation processes. We do not remove dust artificially (from the PBL or by rescaling the vertical distribution) if the simulated opacity exceeds the prescribed opacity from the DGDM. This is because we want to realistically represent the...
processes of dust lifting and the pathways of dust transport and not include sinks of dust other than those (sedimentation and downward transport) that are explicitly represented in the model. We believe this is one of the best strategies for dust lifting and provides a proper insight into the exchanges of dust between reservoirs and the evolution of thermal tides, the vertical dust distribution, and surface stress.

3.3.3. Dust Tagging

In order to better understand the pathways of dust and better characterize the transfer of surface dust between different reservoirs during the GDS, we used the tagging method to track dust in the reference simulation according to its location of lifting. Figure 2 shows the different regions that we consider in this paper. These regions correspond to the most active dust lifting centers during the MY34 GDS. We also tagged dust injected into the atmosphere from the intense lifting occurring in a small area in the Tharsis region around L_s = 198° during the mature stage of the storm. This allows us to follow how and where dust is transported in the atmosphere during the mature phase of the GDS.

4. Model Results: Validation and Overview of the MY34 GDS Phases

In this section, we compare opacities and temperatures simulated by the GCM guided by the MCS observations, discuss the discrepancies, and then present an overview of our best-case reference GCM simulation of the MY34 GDS.

4.1. Validation: Opacities and Temperatures in the GCM Versus Observations

Figure 3 compares the zonal mean observed column dust visible opacities (from the DGDM, see section 2.1) with those obtained with our reference GCM simulation. The simulated opacities are in reasonable agreement with the observations and capture the abrupt increase of opacity around L_s = 187°–188° and a peak of opacity around L_s = 210°. However, simulated opacities are slightly lower than the observations during the onset of the storm and larger during the decay phase of the storm (with opacity differences of up to 0.2).

This is also shown in Figure 4, which compares observed and simulated maps of column dust visible opacity during the GDS. The model captures the main regional storm in Acidalia/Chryse/Arabia during storm onset, the large opacities in the Tharsis/Aonia regions during the mature phase of the GDS, and opacities comparable with observations at all latitudes. However, simulated opacities around L_s = 189°–195° locally peak at ~4 while the observed opacities peak at ~5 during the same period and at the same locations. At L_s = 207°, during the mature phase of the storm, the simulated opacities locally peak at ~9 while the observed opacities peak at ~10 in the Tharsis region and Sabaea Terra but the global mean opacity is larger in the simulation (see section 4.2). These discrepancies are reflected in the $T_{15}$ brightness temperatures, as shown in Figure 5. Simulated daytime and nighttime $T_{15}$ temperatures are globally in good agreement with the observed MCS $T_{15}$ temperatures during the storm. Locally, the agreement is good during the onset of the storm, but we note that the simulated $T_{15}$ temperatures are ~10 K warmer during the peak activity of the storm around L_s = 201°–207° and during the decay phase of the storm. Note that $T_{15}$ temperatures are also sensitive to the vertical distribution of dust, which is not constrained here.
The thermal tides are a global-scale atmospheric response to the diurnally varying component of thermal forcing resulting from aerosol heating within the atmosphere and radiative and convective heat exchange with the surface. The diurnal range of air temperature is a relatively direct measure of dust heating in the atmosphere, and the “diurnal tide” based on $T_{15\_tide} = (T_{3pm} - T_{3am})/2$ is an indicator of where and when heating due to dust absorbing solar radiation is strongest (Wilson & Richardson, 2000).

Figure 6 shows a comparison of observed and simulated 3 am and 3 pm tropical ($30°S$–$30°N$) $T_{15}$ brightness temperatures over the course of the storm. The simulated $T_{15}$ rises quite quickly at storm onset, though not quite as quickly as that observed. This is consistent with the dust opacity somewhat lagging the “observed” opacity, as discussed in section 4.2. It should be emphasized that while the reference simulation has been set up to follow the dust lifting implied by the DGDMs, the vertical distribution and resulting atmospheric temperature structure in the simulation are allowed to self-consistently evolve with the model’s radiative transfer and circulation. Here, $T_{15}$ is dependent on the realistic simulation of the vertical transport of dust by the simulated circulation and is a reasonable representation of the radiative impact of the aerosol.

The zonal wave 1 ($m = 1$) component of the field $T_{15\_tide}$ captures the large-scale zonal variation in the dust heating. Figure 7 shows a comparison of the zonal wave 1 component of $T_{15\_tide}$ at $20°S$ with that observed...
by MCS. Figure 7a suggests enhanced heating in the east at the start of the storm, which is consistent with the initiating flushing storm from Chryse crossing the equator at ~320°E longitude at Ls ~ 186° and subsequently lofting dust that projects onto zonal wave 1. There is a continued migration of the heating center to the east during the growth phase of the storm, with another jump in heating starting at Ls = 196°–197°. We see a plateau in the $T_{15}$ temperature (Figure 6) between these two episodes (Ls = 192°–195°). After Ls = 203°, the storm injection would seem to flatten out, and the zonal wave 1 temperature structure begins to relax as the dust becomes more zonally uniform. By Ls = 215°, the storm is well in the decay phase. The different stages of the storm are well reproduced by our GCM simulation.

4.2. Possible Reasons for Discrepancies Between Model and Observations

The differences between the simulated and observed column dust opacities during the onset of the GDS (i.e., slightly lower in the simulations than in the DGDM) are due in part to an overly large lifted dust effective radius (3 μm) used in the GCM. Larger particles settle rapidly to the surface shortly after being lifted from
the surface, which results in a slower increase in atmospheric dust than what is needed to match the prescribed dust opacity. Using lifted dust distributions with smaller particles help resolve this issue. However, simulations with smaller lifted dust effective radii result in a much longer decay phase of the GDS (as smaller particles settle out more slowly), which is less consistent with observations. This is illustrated by Figure 8. In our reference simulation, the selected effective radius of the lifted dust distribution (3 μm) is therefore the best choice to produce (1) reasonably strong lifting during the onset and development of the GDS, (2) realistic time scales for the sedimentation of the airborne dust during the decay phase of the GDS, and (3) atmospheric $T_{15}$ temperatures in agreement with MCS observations. This choice has consequences on the strength of the Hadley circulation, as discussed in sections 4.3 and 6.1.

In reality, it is possible that small particles are lifted during the GDS onset and then larger particles are progressively lifted as the storm develops and surface wind stresses increase. This could explain why the GCM simulation underestimates the column dust opacity during the onset of the storm and overestimates it during the decay phase. Alternatively, some surface dust reservoirs could contain dust particles of different sizes, thus significantly changing the dust size distribution in the atmosphere over the course of the GDS as dust lifting becomes active in these reservoirs. In the future, using a multimodal dust distribution and adapting the lifting schemes (e.g., to account for different dust particle size distributions depending on the wind stress magnitude or type of surface reservoir) could be of interest to investigate these processes and improve the simulations.

**Figure 6.** The evolution of observed and simulated zonally averaged tropical $T_{15}$ over the course of the MY34 dust storm. Daytime (3 pm) and nighttime (3 am) temperatures have each been averaged over the 30°S–30°N latitude range.

**Figure 7.** (a) Heating anomaly during the MY34 GDS shown as the zonal wave 1 component of the tide field $(T_{3pm} - T_{3am})/2$ at 20°S based on MCS $T_{15}$ observations binned at 2° of Ls. (b) The corresponding tide field based on $T_{15}$ as simulated in our reference simulation.
During the active phase of the storm, the modeled opacity is generally too low in regions of observed dust maxima and too high in regions of observed dust minima. There are several possible reasons that can explain these discrepancies:

1. A "time lag" is naturally produced in the simulation because the MCS-derived DGDMs are daily maps interpolated in time in the model, making it difficult to reproduce a strong daytime or strong nighttime lifting event. In general, the simulation can match a rapid opacity increase but with a time lag of up to 1 sol.

2. The simulation cannot capture high spatial resolution information because the dust maps are available at lower resolution than that of the model.

3. The vertical mixing or the general circulation in the GCM may significantly differ from reality above some regions. For instance, near the southern polar cap edge, local dust lifting in the real atmosphere is presumably caused by strong near-surface winds associated with large temperature gradients. In the GCM, the general circulation continuously transports dust away from the polar cap edge. If this transport is too rapid compared with that on real Mars, this will have resulted in unrealistically strong sources of dust in these regions. Implied cap edge dust lifting may therefore be overestimated in our simulations, which leads to an overestimate of column dust opacities elsewhere, as the regions near the cap edge continuously feed the atmosphere with an excess of dust.

4. The MCS-derived DGDM may contain imperfections. First, the DGDMs are relatively patchy, which is likely unphysical and relates to an observational bias. As the GCM simulation tries to reproduce these opacities, it leads to patchy dust lifting sources and locally underestimated or overestimated opacities. Second, the DGDMs do not capture well the frontal storms that are seen in MARCI images in the high-northern latitudes before onset of the GDS. Other storms occurring near the southern polar cap edge during the GDS may also be missing. This is most likely because these storms are local and rapidly evolving events whereas the DGDM correspond to diurnal averages gridded to ~4° resolution. Additionally, they are shallow dust storms, confined to the lowest atmospheric levels by the descending branch in the high-northern latitudes. These factors make them challenging to detect with MCS limb observations because MCS does not probe the bottom 5 km of the atmosphere where the opacity is too high, especially during storm conditions (Greybush et al., 2019). By contrast, they were detected with TES (a nadir-viewing instrument) and thus captured to some degree in the TES-derived DGDM of MY25. This results in the DGDM produced for TES years versus MCS-only years showing distinct differences in terms of column dust opacities around the cap edges (Montabone et al., 2015). However, frontal storms that do not grow beyond high latitudes may be too short and too shallow to significantly impact the dust cycle, atmospheric temperatures, and the general circulation, so this may not be a key issue for our GCM simulation.

4.3. Sensitivity Studies

We performed several simulations to test the sensitivity of the GCM results to several parameters and modeling approaches in order to try to solve the issues listed above. Changing the effective radius (in steps of 0.5 μm from 0.5 to 5 μm) and the effective variance (in steps of 0.1 from 0.3 to 0.7) of the lifted dust...
distribution improves either the onset or the decay phase of the storm but not both at the same time. Figure 8 shows the variation of opacity with time for a selection of simulations that bracket the observations (black solid line). For instance, injecting 2-μm effective radius particles into the atmosphere (red solid line) allows for better agreement between model results and observations of opacities during the onset phase of the storm but leads to a longer decay phase and larger discrepancies during this period. Using the same particle size but now shutting down the dust lifting in the GCM near the end of the mature phase of the GDS (L\textsubscript{s} \sim 208°, second red dashed line) results in a better agreement between simulated and observed opacities during the decay phase. This suggests that the majority of observed local dust opacity increases during the decay phase were due to transport rather than local lifting, and that perhaps the modeled transport is too slow compared to the rate of dust injection produced by our current scheme.

Injecting dust instantaneously to higher altitudes (up to 25 km) than the computed PBL height slightly reduces the discrepancies during onset but slightly increases them during the mature phase of the storm. In general, changing the vertical mixing in the GCM does not help in solving the issues. We find that if we inject dust into a deep layer, we get a more rapid increase in zonal mean T\textsubscript{15} and in the amplitude of the wave 1 component. The amplitude increase is too rapid if we rescale the dust through the entire column rather than limiting injection to the boundary layer. Removing the highest opacity peaks in the prescribed DGDM (e.g., limiting the maximum column dust visible opacity to 6, 8, or 10) leads to better agreements in opacities and temperatures during the mature phase and the decay phase of the GDS but does not solve the issues during onset and expansion (as lower opacities are prescribed). Allowing dust injection during daytime only or during specific ranges of local times does not significantly change the results. The same applies if we limit cap edge lifting or if we allow dust injection over the polar CO\textsubscript{2} ice caps to match the prescribed column dust opacities in those regions (which is unphysical anyway, as loftable dust is presumably not available there). We also tried a simulation case in which we tune the dust lifting rates (computed in the reference case from the observed DGDM and simulated GCM opacity difference at the specific time and location) by a time-varying factor in order to force the GCM to lift more or less dust (during onset or mature phase of the storm, respectively) than it would do in the reference case. Basically, we used a factor around 1.2 during the onset phase of the GDS and a factor of about 0.5 during the mature phase of the storm so that the simulated GCM dust opacities better match the observed dust opacities. This tuning improves the agreement between the observed and simulated global mean opacities, but significant discrepancies are still obtained locally and overall results remain unchanged. The amount of water vapor and ice clouds and their radiative impact also have little effect on the global evolution of the simulated storm. We also performed a simulation using a 1° × 1° horizontal resolution, but this did not change the results, mostly because the DGDMs are interpolated to a 3.75° × 3.75° grid.

In summary, our results are in generally good agreement with the observations, although we acknowledge small and local discrepancies between the observed and simulated opacities and temperatures. In particular, it has not been possible to match the evolving global mean dust opacity over both the growth and decay phases of the storm. Parts of these issues are attributed to erroneous opacities in the DGDM and to modeling approximations (e.g., related to the way dust is lifted from the surface or vertically transported in the GCM).

However, all of our GCM simulations described in this section show a similar evolution of the GDS, with relatively similar patterns for winds, temperatures, and dust transport, sources, and sinks. This suggests that our GCM results, and in particular those of our reference simulation, are relatively robust.

4.4. Overview of the Different Phases of the MY34 GDS

Figure 9 shows the variation of the zonally averaged dust lifting rate from our reference simulation. The GDS is characterized by zonally averaged dust lifting rates of greater than 0.3 μm per sol (about 7.5 × 10\textsuperscript{−4} kg m\textsuperscript{−2} assuming a density of 2.500 kg m\textsuperscript{−3}), occurring from L\textsubscript{s} = 187° to L\textsubscript{s} = 210° (before the decay phase) and from latitudes 60°S to 60°N. We divide the GDS into four main phases: (1) The onset of the storm, occurring between latitudes 5°S and 60°N and from L\textsubscript{s} = 187° to L\textsubscript{s} = 193°. This phase corresponds to a regional dust storm that develops and moves southwards in Acidalia/Chryse Planitia and Xanthe Terra. Note that significant lifting along the edge of the retreating south polar cap also occurs during this period. The cap edge is simulated to cover latitudes down to at least ±65°S throughout the period of the GDS, thus all dust lifting associated with the cap edge storms occurs at higher latitudes. (2) A rapid eastward and southward expansion of the storm occurring below the equator from L\textsubscript{s} = 193° to L\textsubscript{s} = 196°. During this period, the
regional storm turns into a global storm, triggering dust lifting at all longitudes. (3) A period of intense and maximum dust lifting occurring from Ls = 196° to Ls = 204° between latitudes 20°S and 15°N in the Tharsis/Thaumasia plateau region. Intense dust lifting during this period is also simulated around 60°S, in Sabaea region, Aonia Terra, and along the cap edge. Significantly, more dust lifting within the tropics (15°S–15°N) is involved during this phase of the storm expansion, in comparison with the previous phases. This is also well observed in the T_{\text{15_tide}} field, as shown by Figure 7a, with strong ramping up of the dust heating at Ls ∼ 196° following a brief lull prior to this. (4) The mature phase of the storm, where maximum atmospheric dust loading and maximum atmospheric temperatures occur.

During the prestorm period, dust lifting rates are within the range 0.01–0.1 µm per sol at all latitudes, except near the southern polar cap where the dust lifting rate peaks at about 1 µm per sol. During the decay phase of the storm, little to no dust is lifted in the tropics. This is related to the main issue of the simulation, discussed in section 4.2 and 4.3: the model is unable to reproduce a reasonable decrease in dust opacity during the decay phase of the GDS and tends to simulate opacities significantly larger than those observed. Note that despite simulating too much dust in global average in the atmosphere during the decay phase, the model still produces some dust lifting during this phase, mostly because a few regions clear more rapidly than was observed.

5. Model Results: Evolution of the MY34 GDS

In this section, we explore in detail the simulated sources, sinks, and pathways of dust during the GDS during each of the four phases identified in section 4.4.

5.1. Dust Transport: Hadley Circulation and Thermal Tides
5.1.1. Hadley Circulation

Figures 10 and 11 show the zonal mean atmospheric temperatures with zonal winds and the zonal mean dust mass mixing ratio with mass stream function during the storm, while Figure 12 show maps of diurnally averaged visible dust opacity as obtained in our simulation. During onset of the storm (Ls ∼ 186°–193°), the equatorward low-level return branches of the northern and southern Hadley cells help in transporting and confining dust lifted in the “Hellas” hemisphere (around Hellas and in Acidalia where the main regional storm is active) to the tropics (Figures 11a and 12a and 12b). The slow mean convergence of both Hadley cells forms a narrow equatorial rising branch (in the zonal mean) that promotes the vertical expansion of the dust, transporting it up to 50 or 60 km (Figures 10a and 10b and 11a).

Figure 13 (top) shows meridional winds at two longitudes during this phase of the storm. The diurnally averaged meridional wind is 10–30 m s^{-1} stronger and has more net upper level poleward transport in the “Hellas” hemisphere (left) than in the “Tharsis hemisphere” (right). In other words, the meridional circulation has a strong zonal variation during the storm’s onset and is mainly localized in the “Hellas” hemisphere. This is due to the presence of atmospheric dust in the “Hellas” hemisphere (in particular between longitudes 0° to 60°E where the main regional storm is active, see Figures 12a and 12b), which warms the atmosphere because of the absorption of solar radiation by the dust particles, thus leading to an intensification of the meridional winds. Over the period covering the onset of the GDS (Ls = 187°–193°), the model simulates an increase of atmospheric temperatures of up to 10–20 K at 40 km (10 Pa), as shown in Figures 11a and 11b.

Once dust particles are transported to high altitudes in the tropical regions of the “Hellas” hemisphere, they are transported eastward by the prograde high-altitude winds (Figures 10a and 10b), while remaining confined within the narrow corridor formed by the converging southern and northern Hadley cells (Figures 11a and 11b). Figure 7 shows the eastward migration of the heating anomaly. Figure 12b shows dust clouds between latitudes 15°S–15°N, in Elysium Planitia and above the Tharsis regions during the period.
Ls = 190°–193°, which correspond to high altitude dust transported eastward through this equatorial corridor. As noted in section 2.2, this prograde equatorial jet (super rotation) is sensitive to eddy motions and diurnal thermal tides. The initial regional storm in the “Hellas” hemisphere creates an asymmetry in the longitudinal distribution of atmospheric dust loading, which reinforces eddy motions and the strength of the prograde jet. This is an important feature of the GDS, as it allows for rapid zonal transport of dust.

During the period of global expansion (Ls = 193°–196°), the Hadley circulation continues to strengthen, in association with stronger equatorial zonal winds (dust is still transported eastward through the narrow equatorial corridor), warmer temperatures, and a larger vertical extent of dust, transported upward up to 70 km altitude (Figures 10c and 11c). Large dust opacities of about 3–5 are simulated between longitudes 60°E and 300°E, part of it corresponding to dust transferred from the “Hellas” hemisphere to the “Tharsis” hemisphere (Figure 12c; see also section 5.3).

By Ls = 200° (the middle of the maximum lifting phase), the Hadley circulation transitions to a single cell structure with the meridional flow dominated by a circulation from the summer to the winter hemisphere, as shown by Figures 11c–11f. The zonal wind in the equatorial regions between 10- to 30-km altitude becomes weaker, while the prograde jet strengthens in the winter (northern) hemisphere and weakens in the summer (southern) hemisphere (Figures 10c and 10d). By Ls = 205° (the mature phase), the zonal circulation above the low-to-mid southern latitudes reverses to a weak westward flow, with a weak eastward flow above the South Pole (Figures 10d and 10e). By Ls = 210°, the Hadley circulation decreases in intensity with the decrease of atmospheric dust loading. The westward flow in the low-to-mid southern latitudes strengthens, while the eastward flow above the South Pole weakens (Figure 10f).

5.1.2. Thermal Tides

The impact of the thermal tides on atmosphere temperatures can be seen in Figure 5, which shows a difference of about 30 K between the daytime and nighttime brightness temperatures. The global warming of the atmosphere (as more dust is injected during the GDS) causes an expansion of the atmosphere and the

![Figure 10. Zonal mean atmospheric temperatures averaged over 4 sols (filled contours) as simulated by our reference simulation at roughly 4 sol intervals from Ls = 188° to Ls = 213°. Zonal mean zonal winds are shown as black contour lines.](image-url)
strengthening of the thermal tides (and of the Hadley circulation), with up to 50 K temperature difference between daytime and nighttime at high southern latitudes during the mature phase of the storm (Figure 5, bottom panel). The thermal tides have a strong impact on the circulation. Whereas nighttime cooling of the atmosphere leads to atmospheric contraction and downward motion in the tropics, daytime heating leads to atmospheric expansion and upward equatorial motion, as shown by Figure 13 (bottom). Note that the tide has a strong opposite phase in the boundary layer circulation (indicated by the small black arrows on Figure 13, bottom). These winds play a significant role in the transport of dust, and the diurnally varying surface stresses likely play a critical role in dust lifting too (Wilson, 2012a, 2012b). Note that during the onset of the GDS, the tides are stronger in the “Hellas” than in the “Tharsis” hemisphere (not shown).

Figure 14 (top) shows pronounced differences in the height and meridional extent of the simulated aerosol field between daytime and nighttime that is due to the advecting influence of the sun-synchronous diurnal-period thermal tide. The maximum southward dust extent occurs around 1800 LT and is in phase with the simulated phase of the diurnal period temperature variation in the southern hemisphere (not shown). Maximum poleward tide amplitude for meridional wind velocity occurs above 45 km around midday at Ls ~ 206° (Figure 14, top right), advecting more dust to higher southern latitudes. Similar variations in dust cloud height and extent are seen in TES observations (outside of global storms, McConnochie et al., 2009) and in MCS limb retrievals during the MY34 storm (Kass et al., 2019).

In order to assess the net diurnal impact of the tides on the dust transport, we reran the reference simulation for four sols with diurnally averaged insolation. Figure 14 (bottom) suggests that the tides have little impact on net dust transport because the daytime expansion of the atmosphere balances the nighttime contraction. Dust is slightly more spread at high altitude in the alternative simulation with diurnally averaged insolation, suggesting that the net vertical transport induced by the tides during the GDS is downward. We note that the Hadley circulation is less intense by a factor of 2 in the alternative simulation.

Figure 11. Zonal mean atmospheric dust mass mixing ratio averaged over 4 sols (filled contours) as simulated by our reference simulation at roughly 4 sol intervals from Ls = 188° to Ls = 213°. Zonal mean stream function (10^8 kg s^-1) is shown as black contour lines, with positive values (solid black lines) indicating clockwise circulation.
Figure 12. Maps of diurnally averaged visible dust opacity at a reference pressure of 610 Pa (filled contours). As simulated by our reference simulation, every 2 sols from $L_s = 187^\circ$ to $L_s = 196^\circ$ (GDS onset and global expansion, panels A–C), and every 3 sols from $L_s = 197^\circ$ to $L_s = 212^\circ$ (maximum dust lifting and mature phase, panels D–F; note the change of color scale). The figure shows the development of the regional storm in Acidalia and its eastward and southward expansion in Arabia, Noachis, Sabaea, and Cimmeria before turning into a global storm with opacity $> 4$ at most longitudes in the tropics and opacity $> 6$ above Tharsis and Aonia. Topography is also shown as black contour lines.
This could be due to a less efficient radiative positive feedback in the atmosphere due to the dust being more spread vertically and potentially absorbing less solar radiation, although this remains to be explored in more detail in future work.

5.1.3. Maximum Dust Lifting and Large Dust Plumes Over Tharsis During Ls = 197°–199°

During the period Ls = 197°–199°, the DGDM shows large localized column opacities in the Tharsis region that require our simulation to inject significant amounts of dust into the PBL. This event is associated with local maximum column visible dust opacities of up to 6 (Figure 12d) and warmer atmospheric temperatures above Tharsis (Figure 5). Figure 15 shows the diurnal evolution of the simulated atmospheric dust mixing ratio over 3 sols during this period for dust tagged as being lifted from Tharsis at Ls = 198° in a region where the dust lifting rate exceeds $3 \times 10^{-7}$ kg m$^{-2}$ s$^{-1}$ (~10 μm per sol, see red circle on Figure 2; note that the observed column opacities used to interpolate are one sol apart, which means that the dust injection

Figure 13. Meridional winds simulated by the reference simulation during the onset of the GDS at Ls = 192°. Top: Meridional winds averaged over all times of day over 4 sols at longitude 30° (left) and longitude 210° (right), showing that the dusty “Hellas” hemisphere (left) produces a stronger meridional circulation than the less dusty “Tharsis” hemisphere does at this time. Bottom: 2 am (left) and 2 pm (right) meridional winds averaged over 4 sols at longitude 30° (corresponding to the longitude of the main regional storm and maximum dust lifting at this time), showing the effect of the diurnal tide. Black arrows emphasize the direction of the flow. Note that the diurnally varying meridional winds at low altitudes (below ~6 km) have the opposite phase of the meridional tide at higher altitudes. The tide component is marked by low-level daytime convergence at the equator and divergence at higher altitudes.
rates remain relatively similar over the course of a sol, at a given location). Following this injection, rapid and dramatic vertical motions transport dust up to 70 km during the daytime (12 pm to 6 pm). Subsequent detached layers of dust are obtained between 20 and 60 km altitude and are transported mostly eastward by the high-altitude winds and westward by the near-surface winds. Figure 15 also shows how atmospheric dust is impacted by the diurnal thermal tides, as it is transported upward and downward over tens of kilometers during the daytime and nighttime, respectively.

The large plumes of dust, obtained in the equatorial regions near Tharsis at Ls ~ 198° and shown on Figure 15, are not transported zonally as fast as they would do if they originated earlier during the previous periods of the GDS because of the transition from eastward to westward winds occurring at low latitudes during this period (Figures 10b–10d). The dust plumes are transported eastward below 40 km altitude due to mid-to-low latitude westerlies at these heights at Ls ~ 198° (Figure 10c) and westward above 40 km due to the low latitude easterlies at these heights.

Note that many plumes are obtained in our simulation, but the plume shown on Figure 15 is the largest obtained during the dust storm (compared to the other plumes, it leads to higher local opacities and

Figure 14. (Top) Simulated zonally averaged 2 am (left) and 2 pm (right) aerosol mass mixing ratio for Ls = 206°, with meridional winds (m s\(^{-1}\)) shown with black contour lines. (Bottom) Zonal mean atmospheric dust mass mixing ratio averaged over 4 sols at Ls = 196°, with zonal mean streamfunction (10\(^8\) kg s\(^{-1}\)) shown with black contour lines. Bottom left comes from the reference simulation (as in Figure 11) while bottom right comes from an alternative simulation (restarted from Ls = 194°) performed with diurnally averaged insolation.
Figure 15. Longitudinal cross section of the atmospheric dust mass mixing ratio averaged over latitudes 45°S to 40°N, during 3 sols within Ls = 197°–200° at six different local times (for longitude 0°, snapshots at 12 pm, 4 pm, 8 pm, 12 am, 4 am, and 8 am). Only the tagged contribution of the intense dust lifting over Tharsis at Ls = 198° is shown (see location on Figure 2) in order to highlight and track the plume of dust subsequent to this lifting.
transports dust at higher altitudes). These plumes are consistent with MCS observations of several dust convective events during the MY34 GDS (Heavens et al., 2019). In particular, the large plume shown on Figure 15 is consistent with the specific MCS observation of dust convective activity at similar times (Ls ~ 198°) and locations (near Tharsis). Note that large dust plumes produced during GDSs were also shown in previous modeling, for example, Wilson and Hamilton (1996).

What triggers the dramatic vertical motions of dust obtained in our simulation, and in particular, that above Tharsis? A possible mechanism is the “solar escalator” (Daerden et al., 2015), in which solar heating of a thick layer of dust confined near the surface drives a strong atmospheric heating anomaly and subsequent rapid vertical motions. This mechanism is at the origin of the formation of “rocket dust storms”, which are deep mesoscale convective events able to inject dust at high altitude (Spiga et al., 2013). Here, the evolution of the dust plumes shown on Figure 15 and resulting from the intense lifting over Tharsis resembles that of the “rocket dust storms” but may originate from a separate mechanism. Note that the dust plumes in the GCM cover up to 60° of longitude and correspond to large-scale events, whereas the rocket dust storms usually refer to mesoscale events. In order to investigate what triggers the vertical transport of dust as seen on the left column of Figure 15 (and if the “solar escalator/rocket dust storm” mechanism is at play here), we ran an alternative simulation in which the intensely lifted dust over Tharsis (once again tagged as described above) is replaced by radiatively passive dust, while the rest of the lifted dust remains radiatively active. In other words, the dust lifted that goes on to form the distribution shown in Figure 15 when active is now treated as passive.

Figure 16 shows that the radiatively passive dust is efficiently transported upward, although it reaches lower altitudes (~50 km) than the active dust on Figure 15 (~70 km). This demonstrates that the mechanism leading to the formation of the dust plumes seen on Figure 15 is mainly dominated by the Hadley circulation rather than by direct solar forcing (which could still play a role but smaller).
5.2. Dust Sources and Sinks

5.2.1. Period of Onset of the GDS (Phase A, Ls ~ 187° to Ls ~ 193°)

Figure 17 shows the net change in surface dust cover for the different phases of the storm due to lifting and sedimentation. During the prestorm period, Ls ~ 150° to Ls ~ 187° (Figure 17a), dust lifting in the reference simulation occurs mainly near the south polar cap from ~45°S to 60°S and to a lesser extent in the northern plains (Acidalia/Utopia Planitia and Vastitas Borealis) and in Tharsis. By Ls ~ 187°, a regional storm forms in Acidalia/Arabia and develops further south through the Acidalia/Chryse topographic channel and east in Arabia, lifting dust in these regions (Figures 12a and 17b).

Figure 18 shows the relative contribution of dust from the 10 spatially tagged regions shown in Figure 2 over the course of the storm. Note that these 10 regions contribute at least 90% of the lifted dust at all phases of the storm (Figure 18a). A large and sudden increase of simulated dust lifting occurs from ~15°S to 30°N in Arabia (Figure 18a). This event is responsible for the abrupt increase in dust opacity (up to 3) and temperature observed by MCS in this region (Figures 12b and 8, respectively).

Figure 17. Net change in surface dust cover as simulated by our reference simulation, occurring during the (a) prestorm period (Ls ~ 150° to Ls ~ 187°), (b) onset phase, (c) expansion phase, (d) maximum lifting phase, (e) mature phase, and (f) decay phase (Ls ~ 210° to Ls ~ 250°).
During Ls ~ 187°–190°, the regional storm moves eastward to Sabaea and to the west of Elysium and southward to the edges of Hellas and to Noachis. Intense lifting in these regions (Figure 18a) produces dust clouds in the “Hellas” hemisphere, for example, in the general Noachis region to the northwest of Hellas and west of Elysium, with peak visible dust opacities of up to 4 (Figure 12b).

Around Ls ~ 191°, small amounts of dust are lifted in the “Tharsis” hemisphere (e.g., Tempe, Tharsis, and Aonia; see Figure 18a), which suggests that the eastward transport of dust from the “Hellas” hemisphere triggered more lifting in those locations by warming the atmosphere, increasing the heating and the surface stress, and/or eventually supplying the surface with dust particles that can be lifted again (the evidence for each option is discussed in sections 5.3 and 6.3).

By Ls ~ 192°, the main regional dust storm slightly shifts towards the southeast and grows to cover the entire Noachis/Hellas/Cimmeria regions, with visible column dust opacities peaking at about 3.5 (Figure 12b). The storm also develops east of Hellas, in Promethei Terra, thus merging with cap edge dust lifting. Dust that has

Figure 18. Contribution from each tagged region of (a) dust lifting rate (i.e., dust sources), (b) dust atmospheric column mass, (c) net surface dust budget from Ls = 180°, and (d) normalized maximum surface stress over 2 sols. All values are shown in global area-weighted average. For convenience, the color scales are the same as those used in Figure 2.
been transported southward in Noachis now fills Hellas, with visible column dust opacities of up to 3.5 above the basin (Figure 12b).

The intense dust lifting obtained in Elysium and seen on Figure 18a is found in the western regions of Elysium only. At the longitudes and times where the equatorial corridor of dust cloud is obtained (see section 5.1.1 and Figure 12b), the model simulates a net but small loss of dust (Figures 17b and 17c), thus suggesting that dust lifting in Elysium contributes to form this dust cloud. Figure 19 shows how dust lifted from Arabia, Sabaea, and Elysium contributes to the equatorial dust cloud. About 50% of the dust within the dust cloud comes from Arabia and Sabaea (25% each), as a result of eastward transport of dust from these regions, and 45% comes from Elysium (and 5% from Cimmeria and Noachis, not shown).

Over the period covering the onset of the GDS (Ls ~ 187°–193°), accumulation of dust onto the surface is simulated in Cimmeria, Aonia, and Tempe (and above the polar caps, but this remains true all the time as they correspond to permanent sinks of dust), while significant net loss of dust is simulated in the regions where the main regional storm was active (Acidalia, Arabia, Sabaea, east Elysium, and Noachis) and along the southern polar cap edge (Figures 17b and 18).

5.2.2. Period of Global Expansion (Phase B, Ls ~ 193°–196°)

By Ls ~ 193°, dust clouds extend to most longitudes. However, lifting in the “Hellas” hemisphere still dominates, with the main initial regional storm moving eastward and southward and being mostly active north and west of Hellas (in Noachis, Sabaea, west Elysium, and Cimmeria), as shown in Figures 12b, Figure 17c, and 18a.

This period is marked by a “kink” around Ls ~ 194° seen in the observations of the global mean temperatures and opacities and reproduced by the model (Figures 8 and 18b). This is due to (1) a strong decrease in dust lifting in the “Hellas” hemisphere, in particular in Arabia, west Elysium, Noachis, and Sabaea; and (2) the fact that the main regional storm (now expanded towards east during this period) lifts less dust in the new regions of active lifting (e.g., Cimmeria) than it did in the Arabia or west Elysium regions few sols earlier (Figure 18a). As a result, the dust lifting remains moderate during this period (in the global average) as
shown by the reduction in dust lifting rate around $L_a \sim 194^\circ$–$195^\circ$ on Figure 18a (top), which explains the slower increase in global mean opacity (Figure 18b, top). This “kink” marks the transition between the dust lifting dominating in the “Hellas” hemisphere to dust lifting dominating in the “Tharsis” hemisphere.

Note that the strong decrease in dust lifting obtained in the “Hellas” hemisphere (in particular Arabia, west Elysium, Sabaea, and Noachis) around $L_a \sim 190^\circ$–$195^\circ$ (Figure 18a) suggests either exhaustion of available surface dust in these reservoirs (resulting from intense lifting during the onset of the GDS) or raised thresholds for dust lifting due to surface dust depletion or lower surface stresses in these regions as the main regional storm moves eastward (Figure 17d shows that the maximum surface stress in these regions during this period tends to decrease). This is further discussed in sections 5.3 and 6.

Over the period $L_a \sim 193^\circ$–$196^\circ$, weak net accumulation of dust occurs over a latitudinal band from $\sim0^\circ$ to $30^\circ$ N in the regions of Acidalia, Arabia, Elysium, and Tempe (Figure 17c and Figure 18d). Hellas, Terra Sirenum, Elysium, and Acidalia are regions where dust tends to accumulate over all the remaining time of the GDS.

### 5.2.3. Period of Maximum Dust Lifting and Large Dust Plumes Over Tharsis (Phase C, $L_a \sim 196^\circ$–$204^\circ$)

The period $L_a \sim 196^\circ$–$204^\circ$ corresponds to a maximum in dust lifting in the Tharsis, Tempe, and Aonia regions, associated with a less intense increase in dust lifting in Arabia and Sabaea and secondary maxima in lifting in Elysium (Figure 18a). The dust lifting activity in both eastern and western hemispheres leads to a strong increase in the global mean visible dust opacity, which occurs as rapidly and for as long as during the onset period of the GDS (Figure 8 and Figures 18a and 18b).

Between $L_a \sim 197^\circ$ and $L_a \sim 199^\circ$, particularly intense and short injections of dust occur in the Tharsis regions for a few sols, triggering large plumes of dust above these regions (see section 5.1.3). By $L_a \sim 199^\circ$, dust lifting in the Tharsis regions strongly decreases but still dominates the dust lifting and the atmospheric column mass of dust by $L_a \sim 203^\circ$ (Figures 18a and 18b). Around $L_a \sim 200^\circ$–$201^\circ$ (following this intense dust lifting in Tharsis), a second peak of dust lifting is obtained in Arabia and Sabaea, suggesting a connection between these regions and Tharsis (see discussion on this in sections 5.3 and 6.3).

The end of this period ($L_a \sim 200^\circ$–$204^\circ$) resembles the onset of the storm as dust lifted from Arabia follows similar transport routes to those described in section 5.2.1 and is followed by dust lifting in Sabaea, around Hellas (Tyrrenhena, Hesperia, and Promethei), and in Cimmeria (Figures 12d and 12de and 17d). Overall, during this period of the GDS, net loss of dust occurs in Tharsis, Tempe, Aonia, Arabia, Sabaea, and west Cimmeria (Hesperia and Promethei), as shown by Figures 17d and Figure 18c. Note that cap edge lifting is still significant during this period in the tagged Aonia region (Figure 17d). By contrast, net accumulation of dust occurs in Hellas, Noachis, east Cimmeria (Sirenum and Icaria), and the south polar cap and a latitudinal band between $0^\circ$–$45^\circ$N including the low topographic regions of Elysium, Acidalia, Amazonia, Chryse, and north of Arabia Terra.

### 5.2.4. Mature Phase of the Storm (Phase D, $L_a \sim 204^\circ$–$210^\circ$) and Decay ($L_a \sim 210^\circ$–$250^\circ$)

During the mature phase of the GDS ($L_a \sim 204^\circ$–$210^\circ$), dust lifting is intense in Tharsis, Tempe, and Aonia and east Cimmeria (Figures 17e and 18a–18c). Atmospheric temperatures and dust loading are maximum during this period (Figures 5 and 10e and 10f). Large plumes of dust, comparable to those shown in Figure 15, are still simulated in this region, with local opacities reaching $\sim6.5$ (Figures 12e and 12f). Dust lifting is limited in other regions (Figure 18a). On a global scale, the sources of dust still dominate the sinks (dust particle gravitational sedimentation), which is why the global opacity continues to increase during this period (Figures 8 and Figure 18b). The peak of the global mean dust opacity is reached at $L_a \sim 209^\circ$ (Figure 18b) due to dust lifting weakening in most regions but mostly in Tharsis, Tempe, Aonia, and Cimmeria (Figure 18a), and to a reduced area of dust lifting in general, thus triggering the decay phase of the GDS.

A net loss of dust in Tharsis, Aonia, and Tempe (and to a lesser extent in some regions around Hellas) occurs during this period (Figures 17e and 18c). As during the previous period, net accumulation of dust occurs in Hellas, east Cimmeria (Sirenum, Icaria), Noachis, and the south polar cap and a latitudinal band between $\sim0^\circ$N and $45^\circ$N including the low topographic regions of Elysium, Acidalia, Amazonia, Chryse, and north of Arabia Terra. At $L_a \sim 209^\circ$, the time of maximum global mean opacity (or column dust mass), the total amount of dust in the atmosphere equates to a global mean column depth of about $4 \mu m$ (Figure 18b, top).
During the decay phase of the storm, most of the airborne dust is sourced from previously active lifting centers such as the Tharsis region (Figure 18b). Dust lifting is still significant along a latitudinal band at 45°N and along the south polar cap. Large quantities of dust fall back to the surface, mostly in the southern hemisphere between 60°S and 30°S, and in particular in Tharsis, Elysium, Hellas, and Arabia Terra (Figures 17f and Figure 18c).

5.3. Teleconnections Between Dust Reservoirs

In this section, we describe possible mechanisms for teleconnections between the dust reservoirs in the “Hellas” (e.g., Arabia and Sabaea) and the “Tharsis” (e.g., Tempe, Tharsis, and Aonia) hemispheres.

5.3.1. Possible Mechanisms

What could be the mechanisms of teleconnections between the different dust reservoirs? We identified two types of mechanisms, with or without transport of dust particles.

1. With transport of dust (mechanism #1): First, dust transported from one region and deposited to another could supply the surface with enough dust that will eventually be lifted later (mechanism #1.1: surface dust exhaustion, resupply, and lifting). Second, if not actual exhaustion, then depletion of dust in a region could cause the lifting threshold to increase, making it harder to lift further dust there (Mulholland et al., 2013; Newman & Richardson, 2015; Pankine & Ingersoll, 2004). For instance, dust in such a region could be available but hard to lift because of its properties (e.g., compactness) or because of some terrain features (e.g., rocks), which could hide a certain amount of this dust. The accumulation of freshly and uniformly deposited dust in such a region could then reduce the threshold again, and very rapid local dust lifting could occur after the surface is resupplied, producing dust clouds that generate strong thermal gradients at their edges thus leading to further lifting (mechanism #1.2: surface dust depletion raises threshold, resupply lowers threshold, and lifting of dust resumes). Third, even without the resupply of dust to the surface, the transport of airborne dust above a region could lead to an increase of surface winds in response to a temperature anomaly associated with dust heating, and thus to a local increase in dust lifting if surface dust is available (mechanism #1.3: transport of dust above a region, heating, increase of surface wind stress, and lifting). This mechanism could include stronger winds at the edge of a regional dust cloud, where temperature contrasts are strongest and also feedback involving the large-scale meridional circulation.

2. Without transport of dust (mechanism #2): Without any significant advection of dust, the heating of the atmosphere above one region due to the presence of locally suspended dust particles (e.g., above the “Hellas” hemisphere during storm’s onset) could rapidly project into a global mode (strengthening the overall circulation) that will be seen in the other hemisphere (e.g., the “Tharsis” hemisphere), thus stimulating surface stress increase there instead (see, e.g., Martinez-Alvarado et al. 2009; Newman et al., 2002; Strausberg et al., 2005). If dust is available, this increase of wind stress would lead to an increase in dust lifting (that was already in place) or trigger a new dust lifting as wind stress becomes higher than the local threshold (mechanism #2: distant heating, increase of surface wind stress, increase in dust lifting, or dust lifting resumes). This mechanism may exist on Mars and may play an important role in expanding and maintaining the GDS activity by propagating an increase in surface stress or in dust availability (see discussion in section 6.3). In the following, section we highlight the simulation results that hint at teleconnections between dust reservoirs.

5.3.2. Back and Forth Transfers of Dust Between Arabia/Sabaea and Tharsis

In our simulation, during the onset of the GDS, dust is first lifted from Arabia, Noachis, and Sabaea and then transported eastward (in particular through the equatorial “corridor,” see section 5.1.1). We find that most of this dust preferentially accumulates in the Tharsis region (Figure 20a). This is interesting because our simulation shows intense dust lifting in Tharsis few sols later, after Ls ~ 196° (Figure 18a), thus suggesting a teleconnection between these reservoirs at this period of the GDS. Figure 21a shows that the amounts of dust deposited in Tharsis during the storm’s onset are not enough to keep the net surface dust budget (i.e., the net change in surface dust cover) positive in this region. There is a net loss of dust in Tharsis until the decay phase of the storm. However, at Ls ~ 195°, the net loss of dust in Tharsis is comparable to the deposition of dust in Tharsis from other regions (Figures 18c and 21a). This means that, during the expansion phase of the simulation, for every dust particle deposited in Tharsis (transferred from another region), two are lifted there (in order to match the observed column.
opacities). If no dust was available in Tharsis prior to storm onset and it received all of its dust from elsewhere in the first part of the storm, there would not be sufficient dust available for this required lifting. Hence, this suggests that dust was available in Tharsis before the GDS and has been mobilized either from the supply of near-surface and surface dust (mechanism #1.2) or from a direct increase of surface stress (mechanism #1.3 or #2, seen in the simulation, see Figure 18d).

Around Ls ~ 200°–201° (following the intense dust lifting in Tharsis), a second peak of dust lifting is obtained in Arabia and Sabaea, after the one that occurred during storm onset (Figure 18a). This is interesting because there is not a strong increase in surface stress in these regions (Figure 18d), which suggests instead that both the Arabia and Sabaea reservoirs may have been exhausted or depleted during the storm onset, and then resupplied in dust from another region in order to produce this second peak of dust lifting during the mature phase of the storm (mechanism #1.1 or #1.2). Figure 20b shows that large amounts of dust lifted from the Tharsis region during this period of the GDS accumulate in Arabia, Sabaea, and their surroundings, thus supporting the hypothesis of a teleconnection between these reservoirs at this period of the GDS. Figure 21b shows that Tharsis (along with Tempe, Cimmeria, and Noachis) is a region from which significant amounts of dust are transferred to Arabia before this period of the GDS. However, the net budget of surface dust in Arabia does not increase (Figure 21b): before Ls ~ 200°, it remains constant for few sols as the deposition of dust in Arabia from the other regions equals the dust sources in Arabia and then it decreases during the peak of dust lifting (Figure 18a). This suggests that dust was available in this region before the GDS but that the reservoir has not been exhausted after the first peak of dust lifting during the onset of the GDS, hence supports mechanism #1.2 instead.

In the future, we will explore dust transfers between reservoirs and assess their balance on multiple annual cycles.

6. Discussion

In this section, we summarize what we have learned about the MY34 GDS and discuss our results in light of comparisons with the MY25 GDS, regional storms, and nondusty conditions.
6.1. Hadley Circulation and Thermal Tides

6.1.1. The Important Role of the Zonal Circulation and Thermal Tides in the Evolution of the Storm

Figure 7a highlights an evident eastward expansion of dust lifting during the growth phase of the storm. An eastward equatorial expansion of the $m = 1$ component of $T_1$ was also seen in TES data during the expansion period ($L_s \sim 186–193^\circ$) of the MY25 GDS (Fig. 7 in Guzewich et al., 2014). By contrast, the later A-season regional storms, which occur somewhat later than the MY25 and MY34 events, generally exhibit westward migration.

Our modeling of the MY34 GDS shows that the initial regional storm at the equator creates a strong zonal gradient of opacity, temperature and tides (Figure 13), which reinforces the equatorial equinoctial season eastward jet. The model shows that the eastward jet in the tropical regions efficiently transports dust from the “Hellas” to the “Tharsis” hemisphere (Figures 10a and 10b) and thus reproduces the rapid zonal expansion of the storm during onset (Figure 7b). In particular, we show a significant net transport of dust from the active regions of dust lifting at this time, Arabia and Sabaea, to the Tharsis regions (Figure 20a). This eastward expansion of the storm remains at first relatively confined to the equatorial regions due to the converging lower branches of the northern and southern Hadley cells that sweep dust into the upward circulation at equatorial latitudes (Figures 11a and 11b). As the most active regions of dust lifting are located near the equator during this time, large amounts of dust are transported eastward, creating a “corridor” of high-altitude dust above the equator (Figures 12a and 12b and Figure 19). In addition, the Hadley cells tend to confine dust lifted at higher latitudes near the surface and transport it towards the equatorial region where it is then easily transported vertically to higher altitudes.

In this paper, we also highlighted the strong effect of the thermal tides on diurnal dust transport. Diurnal thermal tide forcing is usually strongest (with highest temperature amplitude) during equinox, and it strengthens when large amounts of atmospheric dust are present in the tropics, as it is the case for the GDS (Wilson, 2012a, 2012b; Barnes, 2017). Here our simulations show that large plumes of dust are transported downward during nighttime and upward during daytime over tens of kilometers by the thermal tides. The net impact of the diurnal tides on dust transport is a small downward transport (but this remains to be explored in detail over all the phases of the GDS, which is out of the scope of this paper).

6.1.2. Sensitivity of the Hadley Circulation to Dust Loading and Dust Properties

The increase in atmospheric dust loading and subsequent heating of the atmosphere lead to a higher thermal contrast between the equatorial and polar regions resulting in a more intense mean meridional circulation, a more extended Hadley circulation, and a positive amplifying feedback as more dust is injected into the atmosphere from increasing surface stress lifting.

This is in agreement with previous modeling of Mars’s atmosphere showing an increase of the intensity of the Hadley circulation with higher dust loading (Basu et al., 2006; Wilson, 2012a). This is highlighted in Figure 22, which shows a direct comparison of the general circulation (mean meridional and zonal circulation, atmospheric temperature, and dust mixing ratio) obtained with the MGCM during the MY34 GDS (reference simulation) and during MY33 at a similar season. The MY33 simulation was performed with same model setup as the reference simulation but with dust lifting guided by MY33 dust maps.

The intensity of the Hadley circulation in our reference simulation is characterized by a meridional mass flux of up to $60 \times 10^6$ kg/s at $\sim$20 km in altitude in both the northern and southern Hadley cells. Sensitivity experiments confirm that the intensity of this circulation is sensitive to the dust particle size distribution, whereas it remains relatively insensitive to the other simulation parameters listed in section 4.3 (e.g., the altitude and time of dust injection and the model grid resolution). The maximum meridional mass flux decreases by a factor of 2 when the reference radius for lifted dust is decreased from 3 to 1.5 $\mu$m. This is because increasing the effective particle size significantly decreases the ratio of dust opacity in the visible compared to the IR, thus making the dust a more efficient cooling agent in the infrared (Murphy et al., 1993). This is a key parameter in the model as it controls the local radiative balance of the atmosphere and the amplifying positive feedback that dust particles have on the general circulation.

Wilson and Hamilton (1996) reported an increase in Hadley circulation intensity with an increase in IR cooling due to the increase in dust loading. Haberle et al. (1997) used an idealized Mars atmosphere model with Newtonian cooling to show that Hadley circulation intensity was inversely proportional to the thermal damping time scale. In general, the Hadley circulation tends to intensify significantly as the season advances towards the
Figure 22. Comparison between MY34 (left) and MY33 (right) GCM simulations at Ls ~ 198°: 4-sol averaged zonal mean of (a) atmospheric dust mass mixing ratio with streamfunction contours, (b) atmospheric temperatures with zonal wind contours, (c) atmospheric water vapor mixing ratio, and (d) water ice mixing ratio.
Solstice (which is also close to perihelion) and the subsolar latitude migrates further south from the equator (Basu et al., 2006; Haberle et al., 2019). Despite this intensification with season (stronger circulations and surface wind stresses), the dust storm clearly decays after Lₙ ~ 210°, pointing out that negative feedbacks must come to the fore to explain the finite lifetime of the GDS. Such negative feedbacks could include a lack of dust in the source regions although other, as yet unknown, mechanisms could be at work.

6.1.3. Impact of the Circulation on Water Vapor and Water Ice Clouds

Although this paper focuses primarily on the evolution of the dust storm and the pathways of dust, we have also investigated the behavior of water vapor and ice particles in our reference simulation. Figures 22c and 22d show the zonal mean distribution of water vapor and water ice in the reference simulation during the GDS expansion and compares the result with that obtained with the same GCM simulation but performed during MY33 (during which no GDS occurred). During the GDS period, because of the higher temperatures, fewer water ice clouds are obtained, and the remaining ice particles form at higher altitudes. As a result, the atmosphere is enriched in water vapor, which is efficiently transported to high altitudes up to 80 km, along with dust particles, as a result of atmospheric heating and expansion. By contrast, under the nonstorm conditions of MY33, water ice clouds form at lower altitudes and confine water vapor lower down, as ice particles sediment. This is consistent with recent observations of the water vapor abundance in the Martian atmosphere during dust storm conditions, which revealed an increase in high-altitude atmospheric water vapor at high-northern latitudes (Fedorova et al., 2018; Heavens et al., 2018; Neary et al., 2019; Vandaele et al., 2019), and with previous numerical modeling of regional dust storms, which demonstrate that water vapor would be transported to higher altitudes, along with dust, by the large-scale upward motions driven by daytime solar heating of the dust layers (Daerden et al., 2015; Spiga et al., 2013). The transport of water vapor to high altitude and the subsequent breaking of the molecules by UV radiation could explain how water, lakes, and rivers gradually disappeared on Mars over billions of years.

6.2. Dust Sources and Sinks: Comparisons With the MY25 GDS

6.2.1. Overview: Main Similarities and Differences Between MY25 and MY34 GDSs

The MY25 and MY34 GDSs have a nearly identical timing of onset (Lₙ ~ 185° for MY25 vs. Lₙ ~ 187° for MY34) and dust storm peak (between Lₙ ~ 205° and 210°) as well as similar declines in opacity and atmospheric temperatures (Figure 1), although it appears that localized dust lifting in Tharsis (Solis/Syria/Sinai Planum) persisted longer in MY25 (out to Lₙ ~ 225°; Cantor, 2007) than in MY34. However, a major difference is that the MY25 GDS storm was initiated in the southern hemisphere, north of Hellas (Hesperia Planum), rather than in the northern hemisphere. During the MY25 storm, dust was transported eastward along the equator, as in MY34, and evidently subsequent lifting was triggered in Tharsis (Daedalia/Sinai/Solis Planum).

6.2.2. Dust Sources and Sinks During the MY34 and MY25 GDS

Figure 23 shows the net change in surface dust cover (or net budget of surface dust) obtained between Lₙ = 180° and Lₙ = 250° and compared to a simulation of the MY25 GDS, performed with the same model.
setup as our reference MY34 simulation but with dust lifting guided by MY25 dust maps. Over the MY34 GDS, net accumulation of surface dust occurs in Hellas, Cimmeria/Sirenum Terra (30–40 μm), as well as in northern Arabia Terra, Elysium/Amazonis, Hesperia, Noachis, and the southern polar cap (10–20 μm). Net loss of dust occurs in Sabaea/Tyrrhena (20–30 μm), in Aonia/Tharsis/Syria, and along the Thaumasia Plateau (60–80 μm), in Xanthe/Lunea (30–40 μm), and along the southern polar cap at latitude 60°S (20–40 μm). The results obtained for the MY25 GDS show similar patterns, with a significant loss of surface dust in the Tharsis/Syria and Sabaea/Tyrrhena regions and along the polar cap edges. This reinforces the general view that both MY25 and MY34 GDS have similar evolution.

There are differences in the details of southern polar cap edge lifting between the MY34 and MY25 simulations. These must be due to differences in the DGDMs used, which are derived from TES and MCS observations for MY25 and MY34, respectively. Some of those differences may not be physically based but may instead be caused by the differences in how those instruments measure the dust column. For instance, near the cap edge, whereas MCS limb observations would tend to underestimate the dust opacity (because dust is confined near the surface), TES nadir observations would tend to overestimate the dust opacity because of the complicating influence of patchy surface frosts and the weak contrast between surface and air temperatures (Montabone et al., 2015).

It would be interesting to compare the maps of net change in surface dust cover obtained with the GCM with maps of observed albedo changes covering the MY34 GDS period. We anticipate that a net decrease in surface albedo would be seen in the Tharsis/Syria/Aonia regions over a period covering the MY34 (Ls = 180°–250°), as large amounts of dust are removed from the surface in these regions, according to the model (Figure 23a). Conversely, we predict that brighter regions around Hellas, in Sirenum, Hesperia, and Elysium/Amazonis would be seen, as these regions are dominated by net dust deposition during the same period.

While maps of albedo changes are not yet available for MY34, a first comparison between model and observations can be made for MY25. The net changes in surface dust cover obtained between Ls = 180° and Ls = 250° from the GCM simulation of the MY25 GDS are shown on Figure 23b. They are partially in agreement with the MY25–MY26 differences in surface albedo as observed by TES, published in Szwast et al. (2006) (see their figs. 3 and 4) and Smith (2004) (see the fig. 14), assuming that these maps largely indicate where dust was transported during the MY25 GDS. In particular, according to these maps, the Martian surface significantly darkened during MY25 in the regions of Tharsis/Thaumasia Plateau, including Sirenum/Sinai. Our simulation of the MY25 GDS is in good agreement with this observation as it shows a significant net loss of dust in these regions during the period of the GDS (Ls = 180°–250°, Figure 23b), although the observed surface darkening is more extended that the simulated net loss of dust. In addition, the MY25 GDS simulation shows significant net dust deposition in Cimmeria/Sirenum, Hellas, Tyrrhena/Hesperia, northern Arabia Terra, and in the northern low-topographic plains below 45°N, which is in agreement with the observed albedo changes in MY25. However, the GCM predicts strong cap edge lifting for MY25, which is not suggested by the albedo maps. In the regions of Xanthe, Solis, and south Arabia Terra, a net dust accumulation is suggested from the observed TES albedo changes during MY25 whereas a net loss of dust is predicted from both simulations of the MY34 and MY25 GDS. We also note a massive net dust loss over Alba Patera in our simulated MY25, which does not appear in the albedo maps of Szwast et al. (2006).

As our simulations suggest similar evolutions for the MY34 and MY25 GDS, we anticipate that the differences in surface albedo that would be observed over MY34 would be, to first order, similar to those observed over MY25.

6.3. Equilibrium Between Surface Stress and Available Dust

Dust lifting and the subsequent formation of dust storms could be controlled locally by an equilibrium between surface stress and available dust. A local increase in surface stress would allow for larger dust injections into the atmosphere (with an amplifying feedback effect as this would trigger even larger surface wind stresses and lifting) but leave less dust available for future lifting. An increase in available surface dust (as a uniform and “easy to lift” cover) would also favor more dust lifting (see the discussion about teleconnections in section 5.3.1). Conversely, a decrease in surface wind stress or in available surface dust could trigger the decay of a regional or GDS. The balance between these two parameters is subtle on Mars.
In this paper, we highlight two main exchanges of dust between the “Hellas” and “Tharsis” hemispheres occurring during the GDS, according to the GCM:

1. First, the rapid zonal transport of dust (Figures 10a and 10b and 12a and 12b), lifted in the tropics of the “Hellas” hemisphere during onset, quickly supplies the surface of the “Tharsis” hemisphere with available dust. An increase of dust lifting in the Tharsis regions around $L_a = 196^\circ$, subsequent to this significant resupply of dust in this region (Figures 18a and 20a), is obtained.

2. Second, during the period $L_a \sim 196^\circ$–204$^\circ$, increased dust lifting is obtained in the Arabia/Sabaea regions (Figure 18a). This is surprising because large amounts of dust have already been lifted from these regions during the onset of the storm, thus suggesting an exhaustion in surface dust (although it may only be a depletion), and because the maximum surface stresses in these regions remain relatively constant after $L_a = 195^\circ$ (although we note that it increases a bit in Sabaea, Figures 18a–18d). Our simulations show a net transfer of surface dust from the Tharsis regions to the Arabia/Sabaea at this time (Figure 20b).

For the first point, we hypothesized that Tharsis has been supplied in significant amounts of surface dust before $L_a = 196^\circ$. Then, such freshly and uniformly deposited dust could have sufficiently lowered the threshold so that dust lifting resumed over Tharsis (see mechanism #1.2 in section 5.3.1). This could have triggered a local increase of surface wind stress and the further lifting of the previously “hard to lift” dust in this region. Alternatively, or in addition, the transport of airborne dust and subsequent heating above Tharsis (mechanism #1.3) or the remote heating of the atmosphere (mechanism #2) could have been sufficient to locally increase the surface wind stress and trigger the intense lifting in this region, as simulated by the GCM.

For the second point, we hypothesized that the increase in dust lifting in the Arabia/Sabaea region after $L_a = 196^\circ$ could be due to a resupply of fresh and uniformly deposited surface dust, which could then have been easily lifted. A small increase in local surface wind stress in Sabaea could have triggered further lifting of previously “hard to lift” dust in this region (see mechanism #2 in section 5.3.1).

Finally, the GDS begins to decay at $L_a \sim 210^\circ$, while the simulated maximum diurnal stresses are still globally increasing (Figure 18d). This suggests again that the availability of surface dust plays a role in ending the storm. In fact, at this time, the area of dust lifting seems to be reduced to small areas in the Aonia, Tharsis, and west Cimmeria regions (Figures 17e and 18c). It is possible that these regions become exhausted or significantly depleted in surface dust around $L_a = 210^\circ$. Alternatively, or in addition to that, the tropical zonal circulation in the lower atmosphere is very weak at this time as eastward flows change over to westward flows in the southern hemisphere and therefore the zonal transport of dust is less efficient than that during the onset of the storm. This may also play a role in triggering the decay of the GDS, as the dust reservoirs outside the Aonia, Tharsis, and west Cimmeria regions may not have been sufficiently replenished (because of the weak zonal circulation) to maintain significant lifting. As a result, dust sedimentation dominates dust lifting and the GDS decays.

These teleconnections between the reservoirs of dust remain difficult to assess in detail with our simulations because the dust lifting (and subsequent heating, stronger circulation, and increase of surface stress) is controlled by the DGDM rather than being able to evolve self-consistently with the circulation. In the future, we plan to look for observed albedo changes on Mars (e.g., prestorm and poststorm maps), for other evidence indicating the presence or the absence of surface dust in the different regions of Mars, and for ways to measure the amounts of available surface dust. These investigations may be key to confirming the hypotheses of this paper. In addition, we will perform multiyear simulations with interactive dust lifting schemes and limited dust reservoirs and use the tagging method to further assess the balances between the different dust reservoirs.

6.4. Discussion on the Development of Global Versus Regional Storms

In this section, we discuss why global storms develop out of some regional storms at specific seasons in specific years and not for most other regional storms.

The MY34 GDS appears to have developed from a flushing storm sequence originating in the northern hemisphere. These frontal storms may have brought significant amounts of dust into the tropics and in particular into Acidalia/Chryse/Arabia, which produced a local feedback and increased the surface stress in these
regions (Figure 18d). The subsequent regional storm could have been sufficiently large to cross some threshold of dust availability, transportability, and surface stress so that it activated lifting centers in the “Hellas” hemisphere (Sabaea, Hesperia, and Noachis). In addition, the timing of this event, occurring near equinox, is favorable for zonal (eastward) transport of dust by the tropical westerlies that are present below 40 km in a relatively narrow seasonal interval around equinox (Figure 10a).

During the Martian years without GDS, regional storms generally occur during the dusty season ($L_s \sim 180^\circ$–$360^\circ$). Some of them are characterized by rapid meridional migrations from high latitudes to equatorial regions through the low topographic channels of Acidalia/Chryse, Utopia/Isidis, or Arcadia/Amazonis. In particular, the A and C season storms are usually flushing through these channels. This may be because, at this time, tropical zonal winds are moderate, and as dust is transported through the tropics, it is not efficiently transported zonally and remains relatively confined in longitude.

We note that GDSs seem to occur either near southern spring equinox (MY12, MY25, MY34, $L_s \sim 180^\circ$–$205^\circ$) or near southern summer solstice (MY1, MY9, MY10, MY12, MY28, $L_s \sim 250^\circ$–$300^\circ$, MY1) and maybe at both within the same year (MY12). Although clear observations of GDS are still relatively few, no confirmed GDS has been observed to begin within the A and C season periods ($L_s = 210^\circ$–$235^\circ$ and $L_s = 310^\circ$–$340^\circ$, respectively).

Here, we hypothesize that the development of the GDSs may be related, in part, to rapid zonal expansion in the tropics (eastward during equinox and westward during solstice), which would perhaps allow for greater amounts of dust to be lifted rapidly than during the presolstice and postsolstice seasons. If the right combination of dust availability and surface stress for intense lifting occurs late after equinox or late after solstice, the zonal circulation within the upward branch of the Hadley cell may be too weak to efficiently transport dust zonally, leading to the more meridional evolution of the presolstice and postsolstice regional storms. However, we note that several regular regional storms typically occur around $L_s \sim 180^\circ$ (e.g., Wang & Richardson, 2015) but do not always turn into GDSs despite the favorable season for strong zonal winds and dust transport. This is still not well understood. It is possible that a large initial regional storm is required at first to trigger a GDS and that the combination of dust availability and surface stress during the Martian years without GDSs did not allow that.

Whereas equinoctial GDSs can originate either from dust lifting in the northern (MY34) or in the southern (MY25) hemisphere (but with subsequent significant dust lifting near the equatorial regions and the rising branch of the Hadley cell), we hypothesize that solstitial GDSs may only originate from dust lifting in the southern summer hemisphere because the descending branch of the Hadley cell in the winter Northern Hemisphere would confine dust near the surface and limit its zonal transport. The observation of a regional storm in Noachis immediately prior to the onset of the MY28 GDS supports this hypothesis, although it appeared to merge with a northern storm in Chryse, which may play an important role in the development of the GDS too (Wang & Richardson, 2015). In the future, we aim at investigating these mechanisms in further details with GCM modeling studies.

7. Summary and Conclusions

In this paper, we simulated the MY34 GDS and carried out a detailed characterization of the atmosphere during the GDS (temperatures, dust loading and opacity, winds, and clouds) with the NASA Ames Mars GCM, which includes a dust lifting scheme forced by MCS-derived dust maps. Our results are in generally good agreement with the available observations of the GDS, including the atmospheric $T_{15}$ temperatures and the column dust opacities. Small discrepancies between our simulated dust opacities and those observed are obtained locally. In particular, the model does not simultaneously provide a good match to the onset and decay rate of the GDS. We performed a sensitivity study of our GDS simulation to the dust lifting distribution (effective particle size and variance), to the impact of water ice clouds, and to the vertical mixing. Overall, none of these simulations improved the agreement with the observations, and no significant changes in dust transport or dust sources and sinks were noted, which demonstrates the robustness of our GCM results (see section 4.3).

We highlight a mechanism that may play a role in the storm’s onset and expansion. We find that the initial regional storm at the equator creates a strong zonal gradient of opacity and temperature, which causes
strong eastward winds at the equator. As a result, dust is efficiently transported to the east during storm’s onset, which may significantly contribute to trigger new dust lifting at all longitudes. Dust is also efficiently and rapidly transported upward by the Hadley circulation, in particular in the equatorial regions. Both the Hadley cell and the thermal tides increase in intensity as more dust is injected into the atmosphere due to a positive feedback involving radiative transfer.

We also highlight significant rapid back and forth transfers of surface dust occurring during the development of the storm between reservoirs located in the “Hellas” (e.g., Arabia and Sabaea) and “Tharsis” hemispheres, which result from the rapid zonal circulation (eastward during onset and westward during the mature stage of the storm) at the latitudes of the upward branch of the circulation, characteristic of this “equinoctial” storm. These exchanges of surface dust could correspond to teleconnections between regions and play an important role in the storm development and expansion. For instance, they could allow for fast replenishment of the surface with available dust or for an increase in local heating and in subsequent surface wind stress and dust lifting. Our results suggest that both the Arabia and Sabaea reservoirs may have been exhausted or depleted during the storm onset and then resupplied in dust from another region, in particular Tharsis. They also suggest that dust was available in Tharsis before the GDS and has been mobilized either via (1) the supply of additional dust to the surface having the effect of lowering the threshold or (2) a direct increase of surface stress (described as mechanisms #1 and #2 in section 5.3.1).

Our simulations also show that gigantic dust plumes occur over Tharsis during the mature phase of the storm. These plumes are able to inject dust at high altitude up to 80 km, as a result of dramatic vertical motions driven by the general circulation (see section 5.3). The model also shows that the water ice cloud condensation level migrates to higher altitudes due to atmospheric warming in response to dust heating, which leads to an enrichment of the upper atmosphere in water vapor. This is consistent with recent observations and could explain how water, lakes, and rivers gradually disappeared on Mars over billions of years.

In our simulations of the storm, the intensity of the Hadley cell is significantly stronger than that in nondusty conditions. We find the intensification to be strongly sensitive to the radiative properties of dust particles, which are tied to the effective radius of the lifted dust particle size distribution. This sensitivity to the dust particle size and the impact of nonuniform particle size distribution for the lifted dust (e.g., taking into account variation in time and space and multimodal size distribution) will be investigated in a future work.

This work is to be continued, in particular by extending the simulations over many Martian years, by carefully comparing different GDS simulations and by investigating the exchanges of dust with interactive or semi-interactive dust lifting schemes and tagging methods.

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