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Surface warming during the 2018/ Mars Year 34 Global Dust Storm

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Key Points:

- Mars’ 2018 Global Dust Storm caused a 0.9 K globally-averaged surface warming, but with local 16 K cooling/19 K warming
- The magnitude of dayside cooling was controlled by atmospheric dust, and nightside warming by surface thermal inertia
- The effects were strongly non-uniform, with high dust loading causing net warming (cooling) over low (high) thermal inertia continents
Abstract

The impact of Mars’ 2018 Global Dust Storm (GDS) on surface and near-surface air temperatures was investigated using an assimilation of Mars Climate Sounder (MCS) observations. Rather than simply resulting in cooling everywhere from solar absorption (average surface radiative flux fell 26 Wm⁻²), the globally-averaged result was a 0.9 K surface warming. These diurnally-averaged surface temperature changes had a novel, highly non-uniform spatial structure, with up to 16 K cooling/19 K warming. Net warming occurred in low thermal inertia (TI) regions, where rapid night-time radiative cooling was compensated by increased longwave emission and scattering. This caused strong nightside warming, outweighing dayside cooling. The reduced surface-air temperature gradient closely coupled surface and air temperatures, even causing local daytime air warming. Results show good agreement with MCS surface temperature retrievals. Comparisons with the 2001 GDS and free-running simulations show that GDS spatial structure is crucial in determining global surface temperature effects.

Plain Language Summary

Martian Global Dust Storms (GDS) are planet-encircling events which fill the atmosphere with a deep layer of mineral dust. During these events the dayside of the planet cools due to the blocking of sunlight, but the nightside warms from dust scattering back surface emissions in the manner of the greenhouse effect. We combined observations of the most recent (2018) GDS from an orbiting instrument, the Mars Climate Sounder, with a Mars climate model to study the storm’s effects on surface and near-surface temperatures. We found that the net effect was actually an increase in global average surface temperatures. The cause was the significant night-time warming of regions with low thermal inertia, which normally cool rapidly at night but are provided an atmospheric dust “blanket” by the storm. The magnitude of warming was enough to compensate for the net cooling over areas with higher thermal inertia. Near-surface air temperatures also rose, as the storm coupled these more closely to the surface. Further simulations showed that these results are valid over various possible storm intensities. The role of thermal inertia suggests that the geographical extent of a GDS, and which regions it covers, plays a significant role in its ultimate effects.

1 Introduction

Dust aerosol is a critical component of Mars’ atmosphere, and has long been known to have significant radiative and dynamical effects through scattering and absorption of radiation (e.g. Gierasch & Goody, 1972; Pollack et al., 1979). Global dust storms (GDS; here, events spanning all longitudes over a wide range of latitudes) are a spectacular example of dust-related phenomena on Mars, occurring every few martian years (MYs) and covering swathes of the planet with a deep dust cloud for months at a time (e.g. Haberle, 1986; Leovy et al., 1973; Zurek, 1982; Zurek & Martin, 1993). These storms have been modelled to have substantial effects on the circulation (e.g. Böttger et al., 2004; Bouger et al., 1997; Haberle et al., 1982; Lewis & Read, 2003) and radiative balance (e.g. Read et al., 2016) of the atmosphere.

One way to describe the degree of dust loading in the atmosphere is by optical depth, defined as the log of the ratio of incident to transmitted intensity of a beam at a certain wavelength (Petty, 2006). In practice, the radiative effects of an atmospheric aerosol also depend on particle radius and its specific scattering/absorption properties. Dust generally has a greater scattering effect on incident sunlight than smoke (which is compositionally different and, on Earth, generally smaller (Friedlander, 2000)), which primarily absorbs in the visible;
smoke therefore has a greater “anti-greenhouse effect”, and it has been famously theorised that a global smoke cloud on Earth would result in drastic surface cooling (“nuclear winter”) (Turco et al., 1984), with soot/smoke used in nuclear winter simulations of single-scattering albedo (SSA; ratio of scattering to extinction at solar wavelengths) of (e.g.) 0.64 (Robock et al., 2007). Soil dust alone has been shown to have surface radiative effects which are highly dependent on the specific SSA used, with higher SSAs (0.97 vs 0.84) causing less shortwave flux reduction at the surface (Shell & Somerville, 2007). Recent work on the properties of martian atmospheric dust, based on observations of the 2007 GDS, estimates a SSA of 0.94 (Wolff et al., 2009); significantly greater than that of soot/smoke. Aerosol properties are critical for determining aerosol radiative effects, and SSA in particular has a large impact on shortwave radiative flux at the surface.

High opacities in a Mars GCM have been shown to decrease surface shortwave flux while increasing longwave emission to the surface, with a net reduction in surface flux of ~70 Wm\(^{-2}\) as averaged over an MY for the unrealistic scenario of a visible-wavelength opacity 5 dust cloud covering the planet for a whole orbital cycle (Read et al., 2016). In situ observations of the 2018 GDS from the Mars Science Laboratory (MSL) showed substantial dayside surface and near-surface cooling due to the reduction in shortwave flux, but also a nightside warming effect (Guzewich et al., 2019); this latter due to enhanced longwave emission and backscattering as a result of the increased aerosol and consequently higher atmospheric temperatures (Martínez et al., 2017). Orbital measurements confirm this (see Validation).

Surface properties have also been shown to be key in determining STs and near-surface air temperatures (ATs). The surface thermal inertia (TI) describes the temperature response of the surface to incident energy flux, and is especially important on Mars given the low atmospheric density and lack of oceans to act as heat reservoirs. Materials with low TI, such as loosely aggregated dust, heat and cool rapidly, while materials with high TI (like bedrock) stay relatively warm at night and cool in the day. Ground temperatures at the MSL site, for example, are driven mostly by the local TI, with lower (higher) TI regions resulting in more (less) extreme minimum and maximum ground temperatures (Martínez et al., 2017). These lower nightside/higher dayside temperatures at low TI regions are due to increased radiative heating on the dayside and rapid radiative cooling on the nightside.

2 Methods

2.1 Model

The Mars Global Circulation Model (MGCM) is a four-dimensional numerical model which is the result of a collaborative effort between the Laboratoire de Météorologie Dyamique, the University of Oxford, the Open University, and the Instituto de Astrofísica de Andalucía (Forget et al., 1999). The version used here contains a spectral dynamical core with a finite-difference scheme in the vertical, and a semi-Lagrangian advection scheme (Lewis et al., 2007). Dust is advected by the MGCM using a two-moment scheme with a log-normal size distribution (of representative mean effective radius ~1 µm) (Madeleine et al., 2011) with total column dust optical depths (CDOD) scaled to match assimilated observations (Text S1). The vertical dust distribution was allowed to evolve freely. Model dust is radiatively active, with radiative properties derived from observational work (Wolff et al., 2006, 2009) (see Figs. S4.5 for uncertainties in SSA). Unless specified, all opacities described in the context of the MGCM are true CDOD at 600 nm. The
MGCM was run using a spectral resolution of T42, corresponding to a spatial resolution of ~3.75° (~215 km at the equator), and with 50 vertical levels at constant pressure / surface pressure, with midpoints ranging from ~5 m to ~105 km above the surface. The water cycle parametrizations were not included in order to isolate the effects of dust; besides, the greatest radiative effects of water occur in the aphelion season (Steele et al., 2014). The MGCM includes a detailed TI map derived from orbital measurements (Putzig et al., 2005).

2.2 Mars Climate Sounder data and assimilation technique

The data assimilation scheme used is a version of the Analysis Correction (AC) scheme, created by the UK Met Office for operational use on Earth (Lorenc et al., 1991) and modified for the martian atmosphere (Lewis et al., 1997; Lewis et al., 2007). Temperature profiles are assimilated in the same manner as previously used for TES (Holmes et al., 2018; Lewis et al., 2007) and Mars Climate Sounder (MCS) (Holmes et al., 2019a; Steele et al., 2014) data, while dust is assimilated spatially in the form of columns (Lewis et al., 2007). See Text S2 for further details.

The two assimilated fields, temperature profiles and dust column products, are from MCS, a limb sounding instrument aboard the Mars Reconnaissance Orbiter (MRO). Temperature and dust retrievals extend to altitudes of ~85 km, with an intrinsic vertical resolution of ~5 km (McCleese et al., 2010), though the MY 34 GDS led profiles to start and end at higher-than-usual altitudes for the GDS period. MRO’s sun-synchronous orbit means observations are made at two local times, 0300 and 1500 in non-polar regions (Kleinböhl et al., 2009). Quality control applied to dust retrievals is described in Text S3 (see also Montabone et al., 2019). Before assimilation into the MGCM, dust opacities are converted from 21.6 microns to 600 nm via a conversion factor of 7.3 (Kleinböhl et al., 2011). The retrievals used are the most recently processed available (Kleinböhl et al., 2017). The retrievals used for the GDS itself are v5.3.2.

2.3 Simulations performed

Two MGCM simulations with data assimilation (“reanalyses”) were performed, for MY 30 and 34, in addition to 15 free-running MGCM simulations; all relevant data is freely available on the ORDO repository (link: Streeter et al., 2019). The reanalyses assimilated MCS 3D temperatures and 2D CDOD. MY 30 was chosen because of its relative lack of major dust activity. A pre-existing reanalysis of the MY 25 GDS was also used (Holmes et al., 2019b), using TES temperatures and column dust (e.g. Montabone et al., 2005). The free-running simulations were made to assimilate artificial dust column data, replicating the start date and rough latitudinal extent (60° S to 40° N) of the 2018 GDS but with prescribed, spatially and temporally uniform CDOD as normalised to the 610 Pa level, ranging from 1 to 15 (the MGCM radiative transfer scheme should be reliable to within ~10% error even at the highest of these (Toon et al., 1989); see Text S4).

3 Results

Fig. 1 displays the diurnally-averaged ST difference and the dayside (1500) and nightside (0300) differences during $L_s=200^\circ-220^\circ$, the peak of the 2018 GDS, with local times chosen to match MCS observations. Differences henceforth are in relation to the MY
Mars’ dayside surface underwent cooling up to 39 K, global average value 14 K, due to dust-induced blocking of incident solar radiation. The areas with the greatest cooling include Chryse, northern Hellas, Argyre, Isidis, and Amazonis (Fig. 1b). These are all low elevation regions, and correlate with high dust loading (Fig. 2a). Mars’ extreme topographic variation means topographic lows have a greater column opacity at the surface than highs, if pressure-normalised opacities are identical. Low topography regions therefore have higher CDOD. The result was greater cooling over low topography, up to 39 K, and less cooling over high topography, such as the southern highlands and the Tharsis plateau, of <5 K. Note that maximum warming/cooling values are a function of MGCM resolution.

Mars’ nightside surface underwent warming of comparable degree to the dayside cooling (Fig. 1c), due to the effect of increased backscattering of longwave emission from the surface. This had a magnitude of up to 42 K, with a globally averaged value of 13 K. In contrast to the dayside effects, nightside warming did not correlate with CDOD. This is because the dominant heating effect during the clear-case martian night is surface cooling: highly efficient in Mars’ thin atmosphere. This cooling rate is driven by surface TI, rather than daytime solar insolation. Therefore, the locations of greatest relative night-time warming caused by enhanced longwave backscattering are determined by surface TI rather than by CDOD. The warming is greatest at the high-topography regions of Tharsis and Elysium Mons, but also over the low-elevation Amazonis and Arabia; all low TI regions (Fig. 3b).

In a globally-averaged sense, the nightside 13 K warming was enough to cancel out the dayside 14 K cooling; however, as the two were controlled by independent factors – TI and CDOD, respectively – the diurnally-averaged effect is not one of exact cancellation. Isidis and the southern highlands show a rough cancellation, but most regions do not (Fig. 1a). While there is a net 5 K cooling over Chryse, the greater effect is a net warming up to 19 K over Amazonis/the low TI continents between approximately 160° E - 50° W and 15° S - 40° N (Amazonis/Tharsis/Elysium), between 0° E - 50° W and 10° S - 40° N (Arabia Terra), and Elysium Mons (Fig. 2b). The global effect of the 2018 GDS was therefore a diurnally-averaged increase in STs, due to the strong nightside warming. This was despite a decrease in the diurnal average flux of 10-50 Wm^-2 over most of the planet’s surface (Fig. 2c).

Globally- and diurnally-averaged ATs displayed a 5.3 K increase (Fig. 3b). Nightside AT warming closely tracked ST warming in being greatest over low TI regions (Figs. 1e & 1f); the maximum nightside warming was 37 K. This is because Mars’ ATs are mostly surface-driven. As the nightside surface is warmer during the GDS, so is the nightside near-surface. Less expected is the dayside near-surface warming over some regions, where the surface is cooler (Fig. 1d). This dayside warming reached up to 31 K over the highest parts of Tharsis. The pattern of dayside warming fell into a latitude band between 10° N and 30° S, corresponding to areas of least dayside surface cooling. This is due to the coupling of STs and ATs caused by dramatically increased absorption of both shortwave and longwave radiation in the atmosphere, a result of the increased dust presence. This, together with the reduced shortwave flux on the surface, significantly reduces the surface-air temperature gradient. Therefore, despite the dayside ST cooling from the GDS, the decreased surface-air temperature gradient meant that dayside ATs could be up to 12 K (30 K over Tharsis) higher than in the clear-case. If GDS STs were higher than clear ATs, therefore, then so were GDS ATs.
A range of CDOD (normalised to 610 Pa) were tested to explore the impact of greater dust loadings (over the same region/season as the 2018 GDS) on STs and ATs (Fig. 3a). Increasing CDOD resulted in increased warming/cooling. However, for CDOD >10 the nightside warming magnitude plateaued, remaining constant at ~25 K due to longwave backscattering reaching its maximum efficiency. By contrast, the dayside cooling magnitude continued to increase with CDOD, albeit at a decreasing rate. This exponential-like decay follows from the definition of optical depth as the log of the ratio of incident to transmitted flux. The result is global surface warming for CDOD 1-11, peaking at 3.9 K for CDOD 3-4; this range includes the 2018 GDS. For opacities >11 net global cooling resulted, reaching 1.8 K for CDOD 15.

Globally averaged ATs showed a similar pattern to STs, albeit shifted warmer. Nightside ATs exhibited the same plateau as nightside STs, due to the close coupling between the two. Dayside ATs peaked at optical depth 2, which was sufficient to reduce the surface-air temperature gradient, coupling STs and ATs and thus causing warming, but sufficiently low that the warming was not outweighed by surface cooling. The diurnally-averaged effect was a globally-averaged increase in ATs for CDOD 1-15, peaking as an increase of 8.5 K at opacity 5.

Lastly, we examined ST and AT variation over the course of an average Sol (Fig. 4) at a low TI (10° N, 30° E) and a high TI (5° N, 100° E) location, with near-identical GDS-induced radiative flux differences. The differences in the diurnal ST cycle (Fig. 4a,b) for the clear case are seen in the substantially greater ST variation at the low TI region, especially the much colder nightside temperatures from more efficient radiative cooling. The minimum ST rises 18 K (195 K to 213 K) in the high TI region, but 40 K (from 156 K to 196 K) in the low TI region. Dayside cooling magnitudes are more similar, with maximum ST falling 27 K (284 K to 257 K) in the high TI region and falling 26 K (301 K to 275 K) in the low TI region. As discussed, the magnitude of dayside cooling depends on CDOD and reduced shortwave flux rather than surface properties. The overall effect of CDOD >2 is to reduce the diurnal amplitude of both STs and ATs, by nightside warming and dayside cooling, and to reduce the surface-air temperature gradient, by the coupling mechanism described above. By CDOD 15, the diurnal ST variation decreases from 89 K to 5 K (high TI) and from 145 K to 17 K (low TI).

A major effect of high CDOD was to dramatically decrease the surface-air temperature difference on the dayside. For MY 34, the peak surface-air temperature contrast is 11 K and 21 K for the HTI and LTI regions respectively, compared to 40 K and 55 K for MY 30. The nightside surface-air peak temperature contrast was also reduced from 10 K to 4 K (HTI) and from 21 K to 11 K (LTI), coupling nightside ATs even more tightly to nightside STs.
4 Validation

The MCS surface temperature retrievals of MY 30 and 34 provide an opportunity for validation. As averaged over $L_S=200-220^\circ$, and with MCS’ two local times, the retrievals show a globally averaged net ST decrease of 2.1 K, compared to a decrease of 0.9 K from the MY 34 reanalysis using the same local times. Nightside warming agrees very well with the reanalysis on both morphology (greatest warming over low TI continents) (Fig. S1; data is presented with a seasonal CO$_2$ cap mask applied. For explanation, see Text S5 and accompanying references: Calvin et al., 2017; de la Torre Juarez et al. [this issue]; Kleinböhl et al. [this issue]; McCleese et al., 2008; Piqueux et al., 2015) and in globally averaged value, with warming of 11.2 K and 9.1 K for the retrievals and reanalysis respectively.

Daytime cooling shows greater disagreement, with globally averaged cooling of 15.2 K and 11 K respectively, as well as some disagreement in spatial distribution (Fig. S1). The retrievals agree on high cooling over Chryse and Hellas, but also show high (30+ K) cooling over the southern highlands and Amazonis/Elysium Planitia not seen in the reanalysis. There are a number of possible explanations. Error in CDOD is possible, especially at the high values involved (Montabone et al., 2015); however the results would imply greater CDOD as inferred from the ST retrievals for the dayside, but also smaller CDOD on the nightside i.e. a greater diurnal dust variation. Another explanation is SSA differences; the observed difference is greater than caused by the uncertainty (Figs. S5,6), but an SSA of 5% difference would be sufficient to cause daytime ST differences of 10+ K (Fig. S7). To cause additional cooling specifically, the SSA would have to be in the “dark” part of the observed SSA dichotomy on Mars, contradicting values derived from the 2007 GDS (Wolff et al., 2009). Additionally, the pattern of extra cooling from lower SSA follows locations of greatest CDOD (Fig. S7), and thus does not replicate the cooling pattern in the retrievals; invoking SSA would therefore require heterogeneity in the dust population. Another possibility is that the MGCM’s particle size scheme (Text S1) under/overestimates particle sizes in particular areas, as with greater lifting occurring during a GDS the particle size structure could be far more heterogeneous than usual (Kahre et al., 2008).

Albedo changes could also play a role: large-scale albedo brightening from dust deposition would cause surface cooling by increasing shortwave reactivity, and if deposition was thin enough this would not necessarily alter TI significantly, explaining the good reanalysis-retrieval agreement in nightside STs. Finally, there is the question of more systematic and not necessarily GDS-induced disagreement. While the reanalysis and retrieval nightside STs show very good agreement, there is a systematic daytime bias even in MY 30, a very clear year, of 12 K, going up to 18 K for MY 34 (Figs. S2,3). Further work is needed to investigate this bias; this may result from MCS limb pointing being affected by topography and affecting surface retrievals, but a full investigation of this is beyond the scope of this work.

Overall, the net ST change shows good morphological agreement with the reanalysis: average warming is seen over low TI continents, average cooling elsewhere. One result of the greater cooling in the retrievals is that the net ST change map displays fewer white regions of little/no ST change; boundaries between areas of net warming/cooling are sharper, showing the important effect of surface TI on the ST response.

TES globally-averaged ST retrievals for the 2001 GDS at $L_S=210^\circ$ showed a peak daytime cooling of 23 K and a peak night cooling of 18 K, corresponding to a net decrease of 2.5 K (Smith, 2004). The MY 25 reanalysis shows, for the same time period, a
dayside cooling of 21 K and a nightside warming of 16 K, also corresponding to a net decrease of 2.5 K (note that while nightside STs from the MY 25 reanalysis agree well with TES retrievals, there is a systematic ~10 K disagreement with dayside STs). Averaged over all local times, the reanalysis shows an average ST change of 0 K.

Radio telescope observations of the 2001 GDS found a globally-averaged daytime surface brightness temperature decrease of ~20 K (Gurwell et al., 2005); consistent with the ST cooling in this study (Fig. 3a) and TES observations (Smith, 2004). Hanel et al. (1972) used IR spectroscopy from the Mariner 9 orbiter to examine STs during and after the 1971-72 GDS; the results support broad dayside cooling and nightside warming, but it is difficult to draw any strong or quantitative conclusions given the limited coverage.

The MSL dataset offers a chance for comparison with in-situ ST measurements of the 2018 GDS. Guzewich et al. (2019) show, over $L_\text{S}=195-205^\circ$, a maximum/minimum ST decrease/increase of 22.8 K/15.1 K, corresponding to a net 3.8 K decrease. The MGCM at the resolution used cannot explicitly resolve Gale Crater, so an analogue location at the same latitude (~37.5° E, 5.625° S) was chosen. The TI was 294 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$, compared to the highest published Gale value of 452 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$, and the average CDOD was 5.3, compared to the MSL-measured 5.5. The maximum/minimum ST decrease/increase was 23.4 K/20 K, corresponding to a net 1.7 K decrease. Dayside cooling agrees well, but the MGCM appears to overestimate nightside warming. This is likely due to a lower model TI than that at MSL, which at the time was the high TI Vera Rubin Ridge (Edwards et al., 2018), and any local topographic effects not resolved by the MGCM. Dayside STs also start diverging after $L_\text{S} \sim 210^\circ$ (Fig. S4); possibly due to albedo increases from dust deposition causing surface cooling (Fonseca et al., 2018); the MGCM uses a static albedo map. The MGCM’s ~250 km footprint makes meaningful comparison with a point source like MSL difficult; a mesoscale model could offer a better comparison.

Another in situ source is Viking Lander 1 (VL1), which recorded meteorological data from two major storms (Ryan & Henry, 1979); in both cases, max/min ATs (~1.3 m altitude) rapidly decreased/increased by ~16 K/~12 K, decreasing on average. Qualitatively, given VL1’s relatively high TI location, this matches expectations; however, without better knowledge of opacities a more rigorous comparison is not possible.

5 Discussion and Conclusions

The MY 34/2018 GDS decreased dayside and increased nightside STs, reducing their diurnal variability. Surprisingly, the diurnally-averaged result was a robust and significant net warming over much of the planet. This warming correlated extremely closely with low TI regions, which in clear conditions experience rapid nightside cooling; these regions warmed even as diurnally-averaged total surface flux decreased, due to significant nightside warming from longwave backscattering, which caused nightside ST increases sufficient to outweigh the dayside cooling. Over regions of higher TI, diurnally-averaged STs decreased or remained roughly constant.

Near-surface air temperatures also showed substantial alteration, driven by the surface temperature changes and the reduced surface-air temperature gradient. Even in the clear-case, heat transport in Mars’ atmosphere is dominated by radiation (Barnes et al., 2017; Wolff et al., 2017). Increased dust loading strongly coupled ATs to STs by dramatically increasing radiative absorption (both shortwave and longwave, including of surface emission) in the bottom layers of the atmosphere while reducing shortwave radiative flux at the surface. This
resulted in increased ATs at night and even on the dayside for regions where GDS-case STs surpassed clear NSTs, i.e. Where the clear-case surface-air temperature contrast is greatest.

Interestingly, the MY 34 reanalysis shows less surface warming than the free-running simulation with the same globally-averaged CDOD (Fig. 3a); however, MY 34 surface cooling matches the free runs very well. This can be explained in terms of GDS geographical structure. The 2018 GDS was not spatially homogenous; the highest CDOD were over high TI regions (Fig. 2a), where the nightside warming effect is least. The MY 25 reanalysis, on the other hand, agrees well with the free runs on nightside warming but has stronger dayside cooling. Again, the explanation is geographical: the MY 25 GDS, as represented in the reanalysis (Fig. 2d), had a greater latitudinal extent than the MY 34 GDS, with $\tau_{\text{vis}} > 1$ between 77° S to 66° N versus 69° S to 47° N. This extra area was predominantly high TI, and therefore contributed a net cooling effect. Note that TES had limited latitudinal coverage, and so the MY 25 reanalysis used is constructed from spatially-kriged observations (Montabone et al., 2015); different GDS decay rates could also potentially affect comparisons. The general conclusion holds, however, that GDS spatial structure is important for its overall radiative effects: specifically, the magnitude of dust loading over low vs high TI areas determines the net ST and AT impacts. The MY 34 GDS also shows noticeable diurnal variation in CDOD (Fig. 3), which comes from the variation in MCS CDOD (Kleinböhl et al, this issue); this results in slightly higher dayside cooling/lower dayside warming than in the diurnally uniform CDOD case. The extent to which this is intrinsic variability and not an artefact of MCS dust profile truncation is unclear (Montabone et al., this issue).

One general caveat is that the MGCM uses a static TI map; surface TI has been shown to vary seasonally by up to 200 J m$^{-1}$ K$^{-1}$ s$^{-1/2}$, and to show day-night variability (Putzig & Mellon, 2007). GDS have also been shown to cause lasting alteration of albedo and surface TI via dust redistribution (Szwast et al., 2006; Fenton et al., 2007). That said, seasonal TI variations are very small over low TI regions, suggesting that net warming over these areas is indeed a robust phenomenon. Nightside STs in the MY 34 reanalysis also agree very well with MCS surface temperature retrievals, suggesting good representation of TI in the MGCM. As noted above though, surface albedo changes may affect representation of dayside STs.

Finally, the nightside warming was more persistent in time than the dayside cooling, which mostly affected peak dayside temperatures. The result was that the warming had an outsized impact on diurnally-averaged temperature changes, with more warming in a true diurnal average than in net changes calculated from just two local times. Simulations with varying opacities suggest that a global surface cooling for a GDS with the same structure as the MY 34 event would require a storm opacity of greater than 11; the actual threshold, however, would depend significantly on the storm’s geographical structure.

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especially like to thank reviewers Jim Murphy and Claire Newman for their insightful comments which have substantially improved this manuscript. Reanalyses and free-running simulations are available on the ORDO repository (https://doi.org/10.21954/ou.rd.7902320.v2); MCS data is publicly available on NASA’s Planetary Data System (https://pds-atmospheres.nmsu.edu/).

References


Figure 1. ST (left) and AT (right) difference between MY 34 and MY 30 for the period $L_S=200^\circ-220^\circ$: (top) diurnally-averaged, (middle) at 1500, and (bottom) at 0300. Solid/dashed contours indicate topography above/below areoid.
Figure 2. For $L_S=200^\circ-220^\circ$; (top left) column dust optical depth in MY 34; (top right) surface TI map used in the MGCM; (middle left) diurnally-averaged total surface radiative flux difference between MY 34 and MY 30; (middle right) difference in column dust optical depth between MY 34 and MY 25; (bottom left) column dust optical depth in MY 25; (bottom right) column dust optical depth in MY 30.
Figure 3. (Left, a) ST and (right, b) AT differences relative to MY 30, globally averaged (area-weighted) over $L_s=200^\circ-220^\circ$ for a range of opacities. Presented are diurnal averages, dayside (1500), and nightside (0300). The MY34 (MY 25) GDS is marked with a cross (three-pointed star). CDOD at 610 Pa are also at the relevant local times, and are averaged over 60$^\circ$ S to 40$^\circ$ N.
Figure 4. Averaged over $L_S=200^\circ$-220° for (left) a high TI region and (right) a low TI region: (top) STs, (middle) ATs, and (bottom) the surface-air temperature difference, over the course of a Sol.