Spring 2021

**Observational Analysis and Modeling of the "B" Regional Dust Storm on Mars**

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OBSERVATIONAL ANALYSIS AND MODELING OF THE “B” REGIONAL DUST STORM ON MARS

A Thesis
Presented to
The Faculty of the Department of Meteorology and Climate Science
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In Partial Fulfillment
of the Requirements for the Degree
Master of Science

by
Courtney M. L. Batterson
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The Designated Thesis Committee Approves the Thesis Titled

OBSERVATIONAL ANALYSIS AND MODELING OF THE “B” REGIONAL DUST STORM ON MARS

by

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APPROVED FOR THE DEPARTMENT OF METEOROLOGY AND CLIMATE SCIENCE

SAN JOSÉ STATE UNIVERSITY

May 2021

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ABSTRACT

OBSERVATIONAL ANALYSIS AND MODELING OF THE “B” REGIONAL DUST STORM ON MARS

by Courtney M. L. Batterson

The annually-recurring, regional B storm on Mars occurs at the highest southern latitudes in years lacking a global dust storm (GDS), and produces warm temperatures (> 200 K) at 50 Pa over the south pole. Observations of the B storm are limited due to the lack of in-situ data in the polar regions of Mars, and reproducing polar phenomena using traditional latitude-longitude grid models is difficult because of the increasingly small grid spacing at the poles. The development of the new NASA Ames Mars Global Climate Model (MGCM), which has a finite-volume dynamical core, a uniform cubed-sphere grid, and several of the physics schemes from the NASA Ames Legacy MGCM, provides an opportunity to simulate the B storm at high resolution on a uniform polar grid. This thesis characterizes the evolution of the annually recurring, regional B storm on Mars using MGS/TES and MRO/MCS observations of temperature and dust retrieved from orbit during seven non-GDS Mars Years (MY24, MY26, and MY29–MY34). We define and describe the growth (L_s = 247°–257°), peak (L_s = 267°), and decay (L_s = 277°–287°) phases of the B storm using these observed fields, and then use our analysis to reproduce the storm with the MGCM. We find that the model predicts that dust plumes develop in the eastern hemisphere during the B storm, and that the ascending dust pattern resembles the solar escalator effect. The pluming is well-defined for ~ 5° of L_s around peak intensity (L_s = 267°) and lofts the dust as high as 5 Pa. The model predicts that dust lifting occurs along the receding CO₂ cap edge during the B storm. Model-predicted surface stresses exceed both the fluid and impact thresholds for the saltation of sand-size particles in various regions around the simulated CO₂ cap edge during the simulated B storm.
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At first, only seen
Closer now, yet still so far
The red rock planet.
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1 Introduction

Martian dust storms have been observed and documented since the early 20th century, but their significance was underappreciated until the first artificial satellite of Mars, Mariner 9, successfully entered Mars’ orbit in the middle of a global dust storm (GDS) in 1971 (Kahre et al., 2017; Zurek, 2017). Photographic and spectroscopic data from Mariner 9 prompted the immediate re-examination of previous dust storm observations, and inspired research and discussion regarding the role of dust in the Martian atmosphere (Gierasch, 1974). One of the studies that emerged from the Mariner 9 mission provided some of the first evidence that airborne dust on Mars was an efficient absorber of solar radiation, and that dust could influence large-scale dynamics by altering the local thermal environment in which it is suspended (Gierasch & Goody, 1972; Kahre et al., 2017). Using data from the first year of the Mariner 9 mission, Gierasch and Goody (1972) were able to show that the temperature profiles retrieved by Mariners 6 and 7 could be mathematically reproduced by including the radiative effects of dust in addition to the radiative and convective properties of an atmosphere comprised primarily of carbon dioxide (CO$_2$) gas in the calculation.

It has since been well-established that the Martian atmosphere is highly sensitive to the abundance and distribution of dust in the atmosphere (Kahre et al., 2017). Dust on Mars is a strong absorber of solar radiation and it is therefore capable of radiatively heating the environment in which it is suspended (Pollack et al., 1979). Dust is also a significant absorber and emitter at infrared (IR) wavelengths, and, at high enough concentrations, airborne dust can radiatively cool the local environment (Pollack, Haberle, Schaeffer, & Lee, 1990). By altering the thermal structure of the atmosphere, airborne dust can influence both local- and large-scale atmospheric dynamics, which has important
implications for the condensation and sublimation of CO₂ in the polar regions (Pollack et al., 1979).

The observational record of airborne dust on Mars has grown significantly in the last fifty years (Kahre et al., 2017). As an abundant aerosol in the Martian atmosphere, dust is responsible for the hazy orange hue of the Martian sky and can obscure the surface from view during major dust events. Additionally, a wide variety of dust storm types have been documented on Mars, including small-scale, local dust storms, dust devils, regional dust storms capable of encircling the planet, and large GDSs capable of enveloping the entire planet.

Today, modern global climate models (GCMs) capture aspects of the annual cycle of background dust loading on Mars and can adequately reproduce the effects of dust devils, regional dust storms, and even GDSs (Bertrand, Wilson, Kahre, Urata, & Kling, 2020; Haberle et al., 2019; Kahre et al., 2017; Wu, Li, Zhang, Li, & Cui, 2020). Despite abundant dust storm observations and advancements in global climate modeling, much remains unknown about the physical mechanisms that control the dust cycle and that facilitate the growth, evolution, and cessation of dust storms on Mars. This thesis addresses part of that knowledge gap through an analysis of the mechanisms that raise dust from the surface (lifting) and that cause airborne dust to ascend to higher altitudes (lofting) in an annually-recurring regional dust storm on Mars known as the “B” storm. In the following section, we introduce the physical characteristics that define timekeeping on Mars and summarize the historical record of “B” storm observations that motivate this work.

1.1 Overview

Both Earth and Mars are terrestrial (i.e. rocky, as opposed to gaseous) planets with relatively thin atmospheres and similar day lengths, seasonal climate variations, and
radiative and convective properties (Haberle, Clancy, Forgét, Smith, & Zurek, 2017). A Martian day, known as a sol, is just under 40 minutes longer than an Earth day. Specifically, Mars completes one rotation around its orbital axis every $\sim 24.66$ hours. Mars follows a highly eccentric orbital path at a greater distance from the Sun than Earth, and it therefore requires a total of 687 Earth days (668.6 sols), nearly two Earth years, to complete one orbit around the Sun.

With an obliquity of 25.19°, which is remarkably close to Earth’s obliquity of $\sim 23.44°$, Mars experiences four seasons in a year (Cantor, James, Caplinger, & Wolff, 2001). Seasons on Mars are described according to the areocentric longitude of the Sun ($L_s$), which indicates the location of Mars in its orbit. The beginning of a new Mars Year (MY) begins at vernal equinox ($L_s = 0°$) and every 90° of $L_s$ indicates the equinoctial or solstitial season. Thus, northern spring equinox occurs at $L_s = 0°$, followed by northern summer solstice at $L_s = 90°$, northern fall equinox at $L_s = 180°$, and northern winter solstice at $L_s = 270°$. Years on Mars are counted according to the convention established by Clancy et al. (2000) who designated MY1 as the year in which the first well-documented global dust storm occurred (Zurek, 2017). By this system, it is $L_s = 267°$ of MY35 at the time of this writing (January, 2021).

1.2 Orbital Observations of Regional-Scale Dust Storms

Before Mariner 9 successfully entered Mars’ orbit in 1971, there were three flyby missions that captured a glimpse of the red planet (National Aeronautics and Space Administration [NASA], 2018). In 1965, Mariner 4 snapped the first close-up photos of the Martian surface, and, in 1969, Mariners 6 and 7 performed spectral observations of the atmosphere over the equatorial and south polar regions. In 1976, Viking 1 became the first successful mission to land on the surface of Mars, and, since then, NASA has launched nine successful campaigns to Mars, all of which have carried a variety of
surface and atmosphere observing instruments (NASA, 2018). One of the notable achievements of these missions is the creation and continued development of a robust climatological record of global atmospheric data that has been recorded almost continuously for the last 10 MYs (∼ 20 Earth Years). Retrieved over the course of two orbiting missions to Mars, the Mars Global Surveyor (MGS), which operated from 1997-2006 (MY24–MY27), and the Mars Reconnaissance Orbiter (MRO), which has been operating since 2006 (MY28), are the orbiters responsible for carrying the instruments that record these observations (NASA, 2018).

The Thermal Emission Spectrometer (TES) on board the MGS spacecraft, and the Mars Climate Sounder (MCS) on board the MRO spacecraft have been sampling global and vertical atmospheric temperature and aerosol concentrations for nearly 11 continuous MYs. From these data, Kass, Kleinböhl, McCleese, Schofield, and Smith (2016) identified three highly repeatable, regional-scale dust storms in the southern hemisphere that develop during years lacking a GDS. Named in the order in which they occur, the so-called “A,” “B,” and “C” (hereafter, A, B, C) regional dust storms are defined by elevated temperatures and dust concentrations in the middle atmosphere. Specifically, zonal mean daytime temperatures at 50 Pa (∼ 25 km) exceed 200 K in the regions associated with high dust concentrations during the storms. The dayside and nightside 50 Pa zonal mean temperatures during the MY31 A, B, and C regional storms are shown in Figure 1, which is from Kass et al. (2016). Figure 2 shows the dayside 50 Pa temperatures for MY24, MY26, and MY29–MY32. In these figures, the A and C storm temperatures peak at 50° S during the southern spring season, \( L_s = 215° \), and the late southern fall season, \( L_s = 320° \), respectively. The B storm is observed in the warm region poleward of 60° S around southern summer solstice (\( L_s = 270° \)).
Figure 1. Observed zonal mean 50 Pa temperatures at (a) 3 PM and (b) 3 AM from $L_s = 180^\circ$–$360^\circ$ in MY31. The A, B, and C storms are labeled accordingly, and the A storm northern hemisphere response is labeled in (a). Reprinted with permission from Kass et al. (2016).

The A and C storms are classic planet-encircling dust storms with similar thermal structures. Both storms cause dynamic warming in the northern hemisphere at $\sim 50^\circ$ N (Kass et al., 2016). Observations of the A and C storms suggest they develop from larger sequences of dust storms, specifically southward-propagating flushing storms that originate in the northern hemisphere. This has prompted much discussion about the nature of the A and C storms in relation to other dust events (Battalio & Wang, 2021; Chow, Chan, & Xiao, 2018; Wang & Richardson, 2015).

According to Kass et al. (2016), the A and C storms are strengthened by the radiative heating of airborne dust which is intensified during the perihelion season ($L_s = 180^\circ$–$360^\circ$) when Mars experiences its closest approach to the sun. Heating is further amplified when the subsolar point is at its southernmost location during the southern summer solstice season ($\sim L_s = 270^\circ$). The increased solar radiation during the southern summer season enhances the Hadley circulation whose rising branch is in the southern hemisphere at the latitude of the A and C storms ($\sim 50^\circ$ S). The sinking branch
of the Hadley circulation is in the opposite hemisphere at $\sim 50^\circ$ N which means that, unlike on Earth, the solsticial Hadley cell on Mars is cross-equatorial. The radiative heating of A and C storm dust within the ascending branch of the Hadley cell strengthens the Hadley circulation and subsequently intensifies the adiabatic warming effects of the descending branch of the Hadley cell in the northern hemisphere. The relationship between the A and C storms and the Hadley circulation is likely the process by which the A and C storms produce dynamic warming in the northern hemisphere (Kass et al., 2016).

The physical and dynamical effects of the A and C storms are largely similar, but the storms differ in strength and duration. Whereas the A storm is the most temporally varying of the regional storms, the C storm is the most temporally consistent. In the 50 Pa temperatures in Figure 2, the 200 K isotherm indicates the A storm occurs as early as $\sim L_s = 215^\circ$ in MY31 and as late as $\sim L_s = 235^\circ$ in MY30, but the C storm, on the other hand, consistently develops within 5° of $\sim L_s = 312^\circ$ every year. The C storm is also the shortest of the regional storms, lasting anywhere from 3°–15° of $L_s$ ($\sim 5$–25 Mars days), and it is often the coolest of the regional storms. In fact, the MY24 and MY30 C storms were so cold that Kass et al. (2016) highlighted the 197 K isotherm to outline the MY24 and MY30 C storms in Figure 2.

In MY29 and MY30, Figure 2 shows that the A storm lingers well into the B storm season and delays the onset of the B storm making it difficult to discern exactly when or where the A storm ends and the B storm begins. While the dynamical response produced by the A storm lingers well after the A storm has concluded, the dynamical response produced by the C storm lasts no longer than the storm itself. In fact, the B and C storms are consistently separated by several degrees of $L_s$ and have not been observed to interfere with one another.
Figure 2. Observed zonal mean 50 Pa dayside temperatures during the dusty season ($L_s = 180^\circ$–$360^\circ$) of non-GDS MYs observed by TES (MY24, MY26) and MCS (MY29–MY32). Contours are labeled in the colorbar. The red contour highlights the 197 K isotherm in the MY24 and MY30 C storms and the 200 K isotherm everywhere else. Reprinted with permission from Kass et al. (2016).

The B storm is often the warmest of the regional storms, capable of maintaining 50 Pa temperatures $> 200$ K longer than the A and C storms (Kass et al., 2016). On average, warm ($> 200$ K) 50 Pa temperatures last at least $30^\circ$ of $L_s$ during the B storm but only
15° of \( L_s \) during the A storm and 3° of \( L_s \) during the C storm (Kass et al., 2016). Unlike the A and C storms, the B storm is located south of the rising branch of the Hadley cell and therefore does not produce a dynamical response in the northern hemisphere.

Confined to the highest southern latitudes (60°–90° S), the B storm initiates after perihelion (\( L_s = 252° \)), reaches peak intensity at \( L_s = 267° \) just before southern summer solstice (\( L_s = 270° \)), and decays for \( \sim 20° \) of \( L_s \) (\( \sim 31 \) Mars days) thereafter. As can be seen in Figure 2, B storm initiation varies by \( < 5° \) of \( L_s \) from year to year but the storm consistently peaks in intensity just before southern summer solstice.

1.3 Problem Statement

To date, the A and C storms have featured more prominently in the literature than the B storm. The A storm has been the focus for exploring spatial and temporal changes in eddy kinetic energy (Battalio & Wang, 2020), characterizing the diurnal variation of dust in the middle and high southern latitudes (Wu et al., 2020), analyzing the dynamics of dust storms in the Hellas basin (Chow et al., 2018), and for studying the development of detached dust layers (Heavens, Kass, Shirley, & Piquex, 2019). The C storm was identified as an important mechanism for water vapor transport between the lower and upper atmospheres in Fedorova et al. (2020), its effects on TES retrievals of water vapor abundance are detailed in Pankine and Tamppari (2019), and its dynamic warming effects have been compared to stratospheric sudden warmings on Earth (Mitchell, Montabone, Thomson, & Read, 2015).

The A and C storms have been moderately well-reproduced by GCMs, but no attempts have been made to reproduce a B storm in a GCM to-date (Chow et al., 2018). The B storm is mentioned in the literature as part of larger descriptive studies of dust activity on Mars but has not been studied otherwise (Battalio & Wang, 2021; Wolkenberg et al., 2018). Studying the B storm is challenging primarily because of its location. As
shown in Figure 3, the south pole is far removed from any lander capable of performing in-situ observations of the storm. Successful NASA missions to Mars have placed instruments in the northern hemisphere and in the tropics, but spacecraft have yet to land in the high southern latitudes south of 15° S. Observations of the B storm are therefore limited to data retrieved from orbiting spacecraft.

Figure 3. Locations of NASA spacecraft on Mars as of March, 2021. Credit: NASA/JPL-Caltech.

The location of the B storm also presents a problem for modeling studies. Simulating the atmosphere at high latitudes is difficult to do using traditional latitude-longitude grids because the meridians converge at the poles and cause numerical errors that require modifying the grid near the pole or applying a polar filter to the data. Historically, these latitude-longitude grid models have employed either a short numerical timestep or an east-west Fourier filter to prevent numerical instabilities from developing during the simulation. However, these solutions are computationally expensive, require more computing time to complete a simulation, and in the case of polar filtering, can introduce additional errors in the transported dust field (Warner, 2011).
Recent advancements in GCMs provide an opportunity to simulate atmospheric phenomena in the polar regions at high resolution on a uniform grid. This is made possible by the cubed-sphere grid in the NASA Ames Mars GCM (MGCM) that is illustrated alongside a latitude-longitude grid in Figure 4. The singularity in the latitude-longitude grid is entirely absent in the cubed-sphere grid, and the horizontal resolution is more uniform across the globe in the cubed-sphere grid. In short, the MGCM is ideal for simulating polar phenomena such as the B storm.

Figure 4. An artistic representation of the 2°x2° cubed sphere grid (left) and the 5°x6° latitude-longitude grid (right) over the north pole of Mars. Credit: Alexandre Kling, MCMC, NASA Ames Research Center.

In this work, we expand the observational analysis of the B storm performed by Kass et al. (2016) through MY33, and then characterize the storm according to the TES and MCS temperature and dust retrievals. Then, we reproduce the B storm in the MGCM using observations of the global distribution of dust from MCS in MY31. Finally, we identify and describe the model-predicted dust lifting and lofting mechanisms involved in the initiation and evolution of the B storm.

1.4 Objectives

The goal of this study is two-fold: first, to fully characterize the observed structure and evolution of the B storm, and second, to identify and describe the mechanisms of dust
lifting and lofting involved in its development. The specific objectives required to accomplish the first goal include: reproducing the analysis of the 50 Pa temperatures performed by Kass et al. (2016) for MY24, MY26, and MY29–MY32; extending the analysis through MY33; and describing the growth, peak, and decay phases of the observed B storm. The specific objectives required to achieve the second goal include: reproducing the B storm in the MGCM at high resolution; defining the relationship between the global mean circulation and the development of the B storm; describing the dust lofting behavior in the model; and identifying the model-predicted dust lifting mechanisms involved in B storm development.

This thesis begins with a comprehensive review of the atmosphere and climate on Mars in Section 2. This section includes a high-level description of the thermal structure and molecular composition of the Martian atmosphere as well as an overview of the planetary boundary layer and the CO$_2$, water, and dust cycles on Mars (Section 2.1). This is followed by a summary of the mean circulation at the equinox and solstice seasons in Section 2.2, and concluded with a comprehensive description of dust storms and the processes responsible for the transportation of dust on Mars in Section 2.3.

Section 3 describes the methods and results of the observational component of this work. It addresses the first goal of the study: to characterize the observed structure and evolution of the B storm. The TES and MCS instruments and their observing strategies are described and compared in Section 3.1. In Section 3.2, we present the 50 Pa temperature and dust fields observed during the MY24, MY26, and MY29–MY33 B storms. Using these fields, we define and describe the growth, peak intensity, and decay phases of the B storm and examine its interannual variability.

Section 4 details the modeling component of this work and addresses the second goal of the study: to identify and define the dust lifting and lofting mechanisms in the B storm.
Section 4.1 opens with a description of the MGCM and the relevant physics used to simulate the B storm, Section 4.2 provides an outline of the model initialization settings, and Section 4.3 presents the results of the simulation. In the model results section, the observed and simulated B storms are compared, the mean state of the simulated atmosphere is described, and the details of two sensitivity simulations that were performed to test specific dust lofting and lifting mechanisms in the model are presented. Finally, we close with a discussion of the behavior of dust in the model and how the model might be improved to capture the observed CO$_2$ ice caps at the south pole in Section 5. In the final section, Section 6, we summarize our research and present our conclusions.
2 Background

In this section, we provide a broad overview of the fundamental components of the Martian atmosphere. We begin in Section 2.1 with a description of the composition and structure of Mars’ atmosphere. Then, we compare the thermal structure of Mars’ atmosphere to Earth’s, and we provide a description of the seasonality of the climate and the relevant nomenclature used to describe it. Finally, we detail the planetary boundary layer and the annual CO$_2$, dust, and water cycles on Mars. Section 2.2 contains an explanation of the mean circulation during the equinoctial and solsticial seasons and Section 2.3 outlines the role of dust in the Martian atmosphere.

2.1 An Overview of the Martian Atmosphere

The first evidence for the thinness of the Martian atmosphere came from the Mariner 4 flyby mission in the summer of 1965 (Leovy, 1977). Spectrographic data from Mariner 4 indicated the surface pressure on Mars hovered between 5-7 millibars, which is just 0.62% of Earth’s surface pressure (1013 hPa), and that the abundance of CO$_2$ in the atmosphere was sufficient to explain these observations. Mariner 4 determined that CO$_2$ is the most abundant gas in Mars’ atmosphere, but a successful mission to the Martian surface was required to obtain exact measurements of its relative abundance. During the Viking missions in the 1970s, NASA successfully landed two probes on the Martian surface and reported that the relative abundances of the major atmospheric constituents on Mars are 96% CO$_2$, 2.5% nitrogen (N$_2$), 1.5% argon (Ar), 0.1% oxygen (O$_2$), and trace amounts of krypton, xenon, and water (Leovy, 1977).

With an atmosphere around 100 times less massive than Earth’s, the Martian surface averages a cool 210 K compared to Earth’s 256 K. At the winter poles, temperatures on Mars are so cold that CO$_2$ condenses out of the atmosphere and accumulates on the surface as ice (Leighton & Murray, 1966). This is the process responsible for the growth
and recession of the seasonal CO₂ ice caps at the north and south poles of Mars (Leighton & Murray, 1966). Since Mars’ atmosphere is ∼96% CO₂, the exchange of CO₂ between the surface and the atmosphere each season causes the mass of Mars’ atmosphere to fluctuate as well. This is what is known as a condensing atmosphere and it is unlike any process observed on Earth. In non-condensing atmospheres like Earth’s, the gases that condense out of the atmosphere are minor constituents, such as water, and their phase changes do not significantly alter the mass of the atmosphere.

Mars experiences larger temperature fluctuations than Earth because little water vapor and ozone are present in its atmosphere and there are no oceans to serve as heat reservoirs on Mars (Zurek, 2017). Ozone and water vapor are significant greenhouse gasses that contribute to maintaining a relatively warm climate on Earth. Water vapor is especially important for storing and re-releasing heat overnight. On Earth, the high thermal inertia of the oceans, which cover more than 70% of the planet, serves as a buffer for the diurnal and seasonal changes in the amount of solar radiation received at the surface (National Oceanic and Atmospheric Administration [NOAA], 2009). Lacking sufficient quantities of either gas, and, in the absence of an ocean, the Martian surface receives largely unattenuated shortwave radiation during the day and loses a significant amount of longwave radiation overnight (Zurek, 2017).

In the vertical, the thermal structure of Mars’ atmosphere is somewhat comparable to Earth’s. The Martian atmosphere is divided into three regions according to its standard temperature profile, simply called the lower, middle, and upper atmospheres (Zurek, 2017). The upper atmosphere of Mars is analogous to Earth’s thermosphere. In this region, temperatures increase dramatically with height. The lower boundary of the upper atmosphere, the mesopause, resides ∼100 km above the surface on both Earth and Mars.
Below the mesopause, temperatures on Earth and Mars generally decrease with height (Zurek, 2017).

Between \( \sim 50-90 \) km, which is Earth’s mesosphere and Mars’ middle atmosphere, temperatures on both planets decrease with height at similar rates. However, there is a temperature inversion between \( \sim 20-50 \) km on Earth that is caused by an abundance of ozone located near the top of the stratosphere. This inversion is absent on Mars because, as previously discussed, the Martian atmosphere has very little ozone.

Unlike on Earth, temperatures in the lowest \( \sim 100 \) km of Mars’ atmosphere are highly dependent on the amount of airborne dust present. Between the surface and \( \sim 100 \) km, temperatures on Mars generally decrease with height (Seiff & Kirk, 1977). However, during the perihelion season when Mars is closest to the sun in its orbit \( (L_{\odot} = 180^\circ - 360^\circ) \), the particularly dusty Martian atmosphere is \( \sim 30-50 \) K warmer than the relatively dust-free Martian atmosphere in the opposite (aphelion) season (Zurek, 2017). During perihelion, temperatures in the lowest \( \sim 20 \) km on Mars can increase to Earth-like levels, resulting in a thermal profile that is nearly isothermal with height throughout the lowest \( \sim 50 \) km of the atmosphere. The effects of the dusty season do not alter Mars’ temperature profile above the mesopause \( (\sim 100 \text{ km}; \text{Zurek, 2017}) \).

2.1.1 The planetary boundary layer.

The planetary boundary layer (PBL) is defined as the region of the atmosphere that interacts directly with the surface (Read et al., 2017). It alone accounts for about half of the mass of the atmosphere and, importantly for our study, the PBL plays a crucial role in the exchange of dust between the surface and the free atmosphere on Mars (Hinson, Smith, & Conrath, 2004). Defining the depth of the PBL is important for diagnosing how easily dust is injected into the free atmosphere, but defining the depth of the Martian PBL at a particular location on Mars is difficult because it varies with latitude and temperature.
over diurnal and seasonal timescales. The limited spatial coverage of in-situ data, the
difficulty of observing the near-surface environment from orbit, and the lack of
observations at the poles are additional obstacles inhibiting the direct measurement of the
PBL depth on Mars (Hinson, Pätzold, Tellmann, Häusler, & Tyler, 2008).

Nevertheless, there is some qualitative information about the depth of the PBL on
Mars that has informed quantitative estimates of the average depth of the PBL. Generally
speaking, the Martian PBL is shallower at the high latitudes and during the night. During
the day, the depth of the Martian PBL is estimated to be roughly $\sim 3$–$6$ km on average
and could be as deep as $\sim 10$ km in areas of strong convection or over high-elevation
topographical features (Haberle, Houben, Hertenstein, & Herdtle, 1993; Hinson et al.,
2008). For context, the lower estimate of the depth of the PBL on Mars ($\sim 3$ km) is more
than three times that of the depth of the PBL on Earth, which is typically $\sim 1$ km (Read et
al., 2017). At the high southern latitudes, turbulent mixing within the PBL is unlikely to
loft dust higher than a few hundred meters (Daerden et al., 2015). This is why one of the
goals of this work is to identify the mechanism lofting dust to 50 Pa ($\sim 25$ km) at the
south pole during the B storm.

2.1.2 The carbon dioxide cycle.

The annual cycles of CO$_2$, water, and dust on Mars are three of the fundamental
components of the Martian climate system. The CO$_2$ cycle is the exchange of CO$_2$
between the polar ice caps and the atmosphere each season. As alluded to in Section 2.1,
the seasonal cycle of CO$_2$ on Mars has profound effects on the atmosphere. When
temperatures dip below the condensation temperature of CO$_2$, which is around $\sim 140$ K
near the surface of Mars at the winter pole, as much as 20% of the CO$_2$ in the atmosphere
condenses onto the surface as ice (Leighton & Murray, 1966; Leovy, 1977). During the
summer season, warmer temperatures cause CO$_2$ to sublimate back into the atmosphere,
and the mass of the Martian atmosphere increases as a result. In-situ measurements from
the Viking, Phoenix, Curiosity, and Insight landers indicate that atmospheric pressure
fluctuates as much as 30% with the seasonal sublimation and deposition of CO$_2$ at the
poles (Haberle et al., 2017).

### 2.1.3 The water cycle.

Water vapor is a minor constituent in the Martian atmosphere, but water ice is a
substantial component of both the perennial ice caps at the poles and the ice clouds in the
atmosphere (Kleinböhl et al., 2009). Water ice condensate is predominantly located above
the dust layer. When present at lower altitudes, it is capable of removing dust from the
atmosphere by condensing onto dust particles that act as ice nuclei (Kleinböhl et al.,
2009). The altitude at which water vapor condenses into ice crystals varies seasonally. At
low latitudes, the water saturation level is more than twice as high during the perihelion
season than in the aphelion season because the global perihelion climate is $\sim 20$ K
warmer than the aphelion climate (Clancy et al., 1996; M. D. Smith, 2002). Consequently,
a band of low-level water ice clouds, the Aphelion Cloud Belt or ACB, develops annually
in the northern hemisphere tropics during northern summer (Clancy et al., 1996; Leovy,
1977). The ACB is a manifestation of the cool aphelion climate and the strong vertical
winds that prevail in the northern hemisphere during the aphelion season, and the ACB is
an important regulator of the cross-equatorial transport of water vapor on Mars (Clancy et
al., 1996; Clancy & Nair, 1996). Cool temperatures lower the water saturation level and
strong vertical winds in the ascending branch of the northern summer Hadley cell aid in
the vertical displacement of air in the north at this season (Clancy et al., 2017).

### 2.1.4 The dust cycle.

Dust is present in the Martian atmosphere year-round but it is more abundant during
the second half of the year known as the dusty season ($L_s = 135^\circ–360^\circ$; Kahre et al.,
The dusty season is defined by a background dust haze that is well-mixed in the lowest scale height of the atmosphere \((H \approx 11 \text{ km})\) and less well-mixed at higher altitudes (Kahre et al., 2017). In-situ observations of visible dust opacity indicate that the typical background dust haze increases from an average visible opacity of \(\sim 0.4\) during the clear season to \(\sim 1.0\) during the dusty season, and that local visible dust opacities can peak as high as \(\sim 2–5\) during GDSs (Kahre et al., 2017). The well-mixed background layer is confined to the PBL which is deeper at lower latitudes than at the poles (P. H. Smith et al., 1997). In the presence of strong convection, e.g., during a dust storm, dust particles can be injected out of the PBL and into the free atmosphere where they can then be lofted as high as \(\sim 60–70\) km in altitude (Clancy et al., 2010). It is during the dusty season that GDSs form and, in years lacking a GDS, that the annually-recurring regional A, B, and C dust storms occur.

### 2.2 The Mean Circulation

The mean circulation refers to the pressure distribution, thermal structure, and prevailing wind patterns that define the state of the atmosphere at a particular time of year (Haberle et al., 2019). Atmospheric temperatures, pressures, and winds are averaged over time periods long enough to mask diurnal variations but short enough to exclude seasonal variations from the mean. For Martian phenomena, the averaging period is typically \(\sim 20–100\) sols (Barnes et al., 2017). Global atmospheric temperatures and winds - both zonal and meridional - can be output directly from the MGCM, but only temperature is measured directly from orbit. There is a substantial lack of wind observations on Mars due in part to the failure of several anemometers on various Martian landers (Barnes et al., 2017). As a result, the “observed” zonal wind is derived from the observed temperature field using the theory of gradient wind balance coupled with the assumption that near-surface winds are negligible (Barnes et al., 2017).
Our understanding of the mean meridional circulation on Mars is informed primarily by model simulations of the Martian atmosphere because the meridional wind is not a balanced wind field that can be calculated from temperature observations (Barnes et al., 2017). However, the diabatic circulation, which is a close approximation of the meridional circulation, can be approximated from radiative heating rates which, in turn, can be calculated from the observed dust and temperature fields (Barnes et al., 2017). On Mars, diabatic heating from airborne dust is a major source of atmospheric warming because dust is the most radiatively active constituent in the atmosphere. The mean meridional circulation is heavily influenced by diabatic heating, eddies, and frictional processes, and it can therefore be approximated by the diabatic circulation (Barnes et al., 2017).

In this section, we summarize the mean circulation on Mars during the southern fall equinox ($L_s = 0^\circ$) and southern summer solstice ($L_s = 270^\circ$) seasons, focusing in particular on the state of the atmosphere during southern summer. The zonal mean temperature and wind fields presented in this section are Oxford MGCM-simulated fields from Lewis (2003). The simulation reasonably reproduces the lower atmosphere (0-35 km) temperatures observed by TES in the first year of the MGS mission (Lewis, 2003). Figures 5 and 6 show seasonal averages of the zonal mean temperature, zonal wind, and mass streamfunction during the southern fall equinox season ($L_s = 0^\circ$). Figures 7 and 8 show the same fields during the southern summer solstice season ($L_s = 270^\circ$).

### 2.2.1 Southern fall equinox.

The zonal mean thermal structure of Mars’ atmosphere at equinox, shown in Figure 5, is fairly Earth-like. Temperatures are warmest over the equator and coldest over the poles, creating an equator-to-pole temperature gradient that produces a westerly jet in each hemisphere and an easterly jet over the equator. The strongest winds in the westerly jets are $> 80$ m s$^{-1}$ at 60°–70° latitude in either hemisphere, and the northern westerly jet is
deeper than the southern westerly jet. Winds maximize between 30–0.05 Pa in the north and 1–0.01 Pa in the south. The easterly jet is located between 1–0.01 Pa over the equator and produces maximum wind speeds of \( \sim 40 \text{ m s}^{-1} \) (Cantor et al., 2001).

![Figure 5](image.png)

**Figure 5.** Simulated seasonally and zonally averaged temperature (color-filled) and zonal wind (contoured) during the southern fall equinox season \( (L_s = 0^\circ–30^\circ) \). The zonal wind is contoured at intervals of 20 m s\(^{-1}\), eastward positive (solid lines), westward negative (dashed lines). Reprinted with permission from Lewis (2003).

The zonal mean mass streamfunction, shown in Figure 6, illustrates that, at equinox, there is a thermally-direct Hadley cell in each hemisphere and they share an ascending branch over the equator. The northern Hadley cell descends over \( \sim 40^\circ \text{ N} \) and the southern Hadley cell, which is slightly larger, descends over \( \sim 60^\circ \text{ S} \). There is also a smaller and weaker thermally-indirect Ferrel cell poleward of either Hadley cell. The Ferrel cells are less Earth-like than the Hadley cells. On Earth, the Ferrel cell is bordered...
by a Hadley cell and a thermally-direct polar cell in both hemispheres, but there is no
evidence of such thermally-direct polar cells on Mars in this simulation.

**Figure 6.** Simulated seasonally and zonally averaged mass streamfunction ($10^9$ kg s$^{-1}$) during the southern fall equinox season ($L_s = 0°–30°$). Clockwise negative (dashed lines), anticlockwise positive (solid lines). Zero is represented by the dotted contour. Reprinted with permission from Lewis (2003).

### 2.2.2 Southern summer solstice.

During the southern summer solstice season, the zonal mean thermal structures on Mars and Earth differ greatly. Figure 7 shows that Martian temperatures between the surface and 10 Pa are coolest over the winter pole and warmest over the summer pole during southern summer solstice. This creates an equator-crossing temperature gradient that stretches from the north pole to the south pole. The pole-to-pole temperature gradient is made possible by the low thermal inertia of the Martian surface and the lack of significant amounts of water, and therefore latent heat processes, on Mars (Haberle et al., 2019). The thermal gradient is largest during southern summer because southern summer solstice ($L_s = 270°$) coincides with perihelion ($L_s = 251°$) and therefore the solar insolation received at the surface is maximized.
Figure 7. As in Figure 5 but for the southern summer solstice season (Ls = 270°–300°). Reprinted with permission from Lewis (2003).

The near-surface thermal gradient in Figure 7 is larger between the equator and the north pole than between the equator and the south pole. This produces a strong westerly jet in the northern (winter) hemisphere and a weak but still dominant easterly jet in the southern (summer) hemisphere (Haberle, Pollack, et al., 1993). According to Figure 7, the average speed in the winter jet in the north is \( \sim 160 \text{ m s}^{-1} \), twice the average speed in the summer jet in the south (\( \sim 80 \text{ m s}^{-1} \)).

A fundamental feature of the thermal structure of the Martian atmosphere at solstice is the poleward-tilting “warm tongue” located near the winter pole at around 60° N between \( \sim 1 \text{ Pa} \) and \( \sim 0.05 \text{ Pa} \) (Figure 7). The warm tongue introduces a weak equator-to-pole temperature gradient in the northern hemisphere some \( \sim 50 \text{ km} \) above the surface, and it
forms as a result of dynamic warming in the descending branch of the Hadley cell. With little to no solar insolation at the winter pole, heat in the winter hemisphere is generated by adiabatic processes such as the warming of descending air in the sinking branch of the Hadley cell (Barnes et al., 2017).

The mean meridional circulation on Mars changes dramatically between the equinox and solstice seasons. The pole-to-pole surface thermal gradient shown in Figure 7 helps create a strong, thermally-direct equator-crossing Hadley cell that can be seen in the mass streamfunction field shown in Figure 8. The dominant Hadley cell has a rising branch in the southern (summer) hemisphere at \( \sim 40^\circ \) S and a sinking branch in the northern (winter) hemisphere at \( \sim 50^\circ \) N (Cantor et al., 2001). North of the Hadley cell, there is a weak, thermally-indirect Ferrel cell driven by the baroclinic eddies that control heat and momentum transfer in the polar regions (Barnes et al., 2017). In the opposite season (\( L_s = 90^\circ \), southern winter solstice), the mean meridional circulation, and therefore the large Hadley circulation, is reversed but slightly weaker because the planet is at aphelion, the orbital period when Mars is farthest from the sun and the solar insolation received at the surface decreases.
The cross-equatorial Hadley circulation that forms during the solstice seasons transports dust, water vapor, and clouds from the summer to the winter hemisphere (Barnes et al., 2017). During southern summer ($L_s = 270^\circ$; Figure 8), the maximum mass flux in the Hadley cell is $\sim 100 \times 10^8$ km s$^{-1}$, almost twice the rate of the flux in the Hadley cell in the opposite season ($L_s = 90^\circ$; Figure 6; Barnes et al., 2017; Haberle, Pollack, et al., 1993). By conservation of angular momentum, upper-level zonal winds intensify as a direct result of the strengthened Hadley circulation (Wilson, 1997). Accordingly, the zonal mean zonal wind is stronger in southern summer (Figure 7) than in southern fall (Figure 5).

The southern summer Hadley circulation is strengthened by the increased vertical distribution of heat in the lower atmosphere, which is caused by the intense solar radiation received at the surface during perihelion and the increase in airborne dust during the dusty season (Barnes et al., 2017; Heavens et al., 2011; Wilson, 1997). Other processes that strengthen the southern summer Hadley cell include the CO$_2$ condensation flow and katabatic flows. The CO$_2$ condensation flow occurs as a result of the sublimation and
deposition of CO$_2$ at the poles each season (Toigo & Richardson, 2002). During southern summer, the growth of the northern seasonal CO$_2$ cap removes CO$_2$ from the atmosphere and causes a mass flux of CO$_2$ toward the north as the atmospheric gas is redistributed to maintain equilibrium. Katabatic winds are nighttime downslope flows of cold, dense air that are enhanced in regions of steep topography (Cantor et al., 2001). Mars’ meridional elevation gradient introduces a preference for northward katabatic flows year-round. In short, the intense solar radiation, the increased airborne dust concentration, the CO$_2$ condensation flow, and katabatic flows all contribute to enhancing the northward flux of mass occurring in the southern summer Hadley cell.

### 2.3 Dust on Mars

Dust and CO$_2$ are the two most radiatively active constituents in Mars’ atmosphere, but dust more efficiently absorbs visible radiation than CO$_2$ and it therefore determines where and how much heating occurs (Cantor et al., 2001). Dust is also an efficient absorber and emitter in the IR which enables dust-absorbed solar radiation to be re-radiated at longer wavelengths preferable for CO$_2$ absorption (Cantor et al., 2001). CO$_2$ is more efficient than dust at absorbing in the IR, and therefore CO$_2$-absorbed IR radiation also contributes to radiative heating in the atmosphere, although it does not contribute as significantly as dust (Haberle & Leovy, 1982).

As discussed in Section 2.2, radiative heating in Mars’ atmosphere is highly influenced by the abundance and distribution of atmospheric dust on Mars, which means the presence of atmospheric dust can alter atmospheric dynamics. For example, wind-stress driven dust lifting is amplified when thermal tides and near-surface winds are strengthened by the radiative heating of dust in the atmosphere (Pollack et al., 1979). Conversely, dust lifted by convective vortices (dust devils) has a stabilizing effect. Dust devils are driven by large thermal gradients in the boundary layer, but these gradients are
dampened in an opaque (i.e. dusty) near-surface environment (Newman, Lewis, & Read, 2002). Whether through direct or indirect effects, the abundance and vertical distribution of airborne dust on Mars intensifies temperature gradients and, as we explore throughout this section, can subsequently alter atmospheric circulations (Kass et al., 2016; Wang & Richardson, 2015).

**2.3.1 Global dust storms.**

The dusty season on Mars is marked by a significant increase in dust storm activity, and it is the only time of year in which GDSs are observed. Three GDSs have been observed in the last 20 Earth years (∼10 MYs), the most recent being the MY34 GDS in 2018, and, prior to that, the MY28 and MY25 GDSs in 2007 and 2001, respectively (Kahre et al., 2017). GDSs enshroud the planet in dust so thick that the surface cannot be viewed from orbit for weeks or months at a time. The most recent GDS in MY34 lasted 110 sols and produced global mean visible opacities of 4 and maximum visible opacities as high as 5–10 (Bertrand et al., 2020).

GDSs are most often observed developing from the convergence of several smaller dust storms and usually initiate in the southern hemisphere (Bertrand et al., 2020). However, there is no single location from which these storms consistently initiate. For example, the MY25 GDS grew from several cap-edge storms in the southern hemisphere while the MY28 GDS initiated in the southern midlatitudes, though there is some ambiguity in the data that suggests a northern hemisphere flushing storm may have contributed to its development (Cantor, 2007; Wang & Richardson, 2015). Until the MY34 dust storm in 2018, GDSs had never been observed unambiguously forming in the northern hemisphere (Bertrand et al., 2020). Ergo, the scientific community is still gathering crucial information about the locations of and the processes involved in GDS development.
The timing of GDS initiation is similarly inconsistent. GDSs have only been observed during southern spring and summer (∼Ls = 180°–360°) likely because solar insolation is maximized at this time of year and the global circulation is strengthened as a result (Haberle, 1986). The observational record indicates GDSs have initiated as early as Ls = 185° and as late as Ls = 270°, and that GDSs tend to last several weeks at a time (Bertrand et al., 2020; Cantor, 2007). The MY25 GDS lasted 60° of Ls (∼98 sols), the MY28 GDS lasted 50° of Ls (∼71 sols), and the MY34 GDS lasted 70° of Ls (∼110 sols; Bertrand et al., 2020; Wang & Richardson, 2015). All GDSs observed during the TES and MCS missions and their associated onset, cessation, and duration times are listed in Table 1.

Table 1

<table>
<thead>
<tr>
<th>MY</th>
<th>Storm</th>
<th>Orbiter</th>
<th>Period (Ls)</th>
<th>Duration Ls Sols</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>26</td>
<td>B</td>
<td>TES</td>
<td>250°–295°</td>
<td>45° 70</td>
<td>Kass et al. (2016)</td>
</tr>
<tr>
<td>27</td>
<td>N/A</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>31</td>
<td>B</td>
<td>MCS</td>
<td>250°–290°</td>
<td>40° 63</td>
<td>Kass et al. (2016)</td>
</tr>
<tr>
<td>32</td>
<td>B</td>
<td>MCS</td>
<td>255°–295°</td>
<td>40° 63</td>
<td>Kass et al. (2016)</td>
</tr>
<tr>
<td>33</td>
<td>B</td>
<td>MCS</td>
<td>250°–290°</td>
<td>40° 63</td>
<td>This work</td>
</tr>
<tr>
<td>34</td>
<td>GDS</td>
<td>MCS</td>
<td>181°–250°</td>
<td>69° 110</td>
<td>Bertrand et al. (2020)</td>
</tr>
</tbody>
</table>

2.3.2 Non-global dust storms.

Non-global dust storms include local- and regional-scale dust storms, and distinguishing between the two is somewhat arbitrary as the literature has yet to converge on a standard definition of either one. Originally, regional storms were classified as those with long axes > 2,000 km, so anything smaller than that was considered a local storm (Martin & Zurek, 1993). Today, dust storms are typically defined by both size and duration. Having noticed the diurnal nature of smaller storms and the days-long evolution
of larger storms, Cantor et al. (2001) defines regional storms as those with surface areas > 1.6x10^6 km^2 lasting longer than 3 sols. Wang and Richardson (2015) use this definition with a slight modification, including slightly smaller storms on the order of 1.0x10^6 km^2 in their classification of regional storms. In Kass et al. (2016), the annually-recurring seasonal A, B, and C storms are classified as regional storms in accordance with the modified definition put forth by Wang and Richardson (2015).

Regional dust storms were some of the first dust storms ever observed on Mars because their size and duration makes them visible from Earth (Cantor et al., 2001). Regional storms occur in both the northern and southern hemispheres and are especially active during late southern spring and southern summer, \( \sim L_s = 130^\circ–250^\circ \) (Cantor et al., 2001; Wang & Richardson, 2015). Regional dust storms can be planet-encircling dust storms, meaning that dust in a regional storm can encompass entire latitude bands, but they cannot cover the globe as that would then be classified as a GDS (Wang & Richardson, 2015). The most common planet-encircling regional storms are circumpolar dust storms which frequently occur in the southern hemisphere (Wang & Richardson, 2015).

Like GDSs, regional storms can grow large enough to affect the global circulation by altering the distribution of heat in the lower atmosphere (Wang & Richardson, 2015). Also like GDSs, regional storms are often formed from the accumulation of smaller dust storms. In fact, the largest regional storms tend to form this way (Cantor et al., 2001; Wang & Richardson, 2015). For example, the largest regional storm documented by Cantor et al. (2001) formed when three regional-size storms merged into one massive storm covering an area of more than nine million square kilometers (\( \sim 9.1x10^6 \text{ km}^2 \)).

Local storms are small, on the order of hundreds of kilometers in area, and their development can be initialized by cap-edge winds, local thermal inertia variations,
topography, and baroclinic waves (Cantor et al., 2001). Local dust storms are abundant in the Martian atmosphere. Of the 783 dust storms characterized in the statistical analysis by Cantor et al. (2001), 771 were local storms and just 12 were regional storms. The smallest local storm documented by Cantor et al. (2001) covered an area of just \( \sim 560 \text{ km}^2 \). Local dust storms frequently form around the north and south polar cap edges and also tend to form near the subsolar point in the midlatitudes (Cantor et al., 2001).

2.3.3 Dust lifting.

There are three processes by which particles leave the surface and move through the atmospheres of Earth-like planets, moons, and other celestial bodies. These processes are called creep, saltation, and suspension. Particles in creep are too large and too heavy to be lifted off the surface, so they are pushed forward by the wind or other saltating grains (Greeley & Iversen, 1985). During saltation, sand-sized particles are mobilized and, too heavy to enter into suspension, they fall back to the surface where they either bounce into the air again or become trapped among other particles. Dust in suspension is carried long distances by the wind and can be initiated by saltation or by the wind itself (Greeley & Iversen, 1985). Whether particles move in creep, saltation, or suspension depends on their size. Large particles that move in creep have diameters \( \geq 2,000 \mu\text{m} (\sim 2 \text{ mm}) \), sand-size particles that move in saltation are hundreds of microns in diameter, and fine particles that enter directly into suspension have diameters \( \leq 10 \mu\text{m} \) (Greeley & Iversen, 1985).

Particles are mobilized when the stress imparted by the wind on the surface exceeds a size-specific threshold value (Greeley et al., 1994; Sagan & Pollack, 1967). On Mars, as on Earth, sand-size particles on the order of \( \sim 100-200 \mu\text{m} \) in diameter are most easily mobilized because particles of this size are not so massive that they cannot be lifted off the surface but they are large enough to overcome the strong inter-particle cohesive forces that hold small particles together (Gierasch, 1974; Greeley et al., 1994; Hess, 1973; Kahre
et al., 2017; Sagan & Pollack, 1967). The surface stress threshold for sand-size particles on Mars has been estimated to be anywhere from 10–50 mN m$^{-2}$, but recent estimates tend to lean toward the lower end of the spectrum (Kahre, Murphy, & Haberle, 2006; Newman et al., 2002; Swann, Sherman, & Ewing, 2020). In the NASA Ames Legacy MGCM, for example, a surface stress threshold of 22.5 mN m$^{-2}$ yielded lifting patterns comparable to the geographical lifting patterns observed by Cantor et al. (2001) in both Haberle, Murphy, and Schaeffer (2003) and Kahre et al. (2006).

Surface stress, $\tau$, is a quantity derived from frictional velocity, $u^*$, which is estimated using wind tunnel observations. Surface stress and frictional velocity are related by

$$\tau = \rho u^*^2$$  \hspace{1cm} (1)

where $\rho$ is the near-surface air density (0.02 kg m$^{-3}$; Swann et al., 2020). The wind velocity, $u$, at altitude $z$ can also be derived from frictional velocity using

$$u(z) = \frac{u^*}{k} \ln \frac{z}{z_0}$$  \hspace{1cm} (2)

where $k$ is the von Karman constant (0.40) and $z_0$ is the altitude where the Law of the Wall predicts zero wind velocity (0.0001 m from Swann et al., 2020). Thus, the experimentally-derived frictional velocity for saltation can be converted to 1.5 m wind velocities for comparison with in-situ observations of the Martian wind. Additionally, the relationship between frictional velocity and surface stress in Equation 1 is useful for comparing threshold values across the literature.

The threshold frictional velocity for the saltation of sand-size particles ranges from $\sim 1.5–2.5$ m s$^{-1}$ in a 10 mb atmosphere, to 4–5 m s$^{-1}$ in a 5 mb atmosphere (Greeley & Iversen, 1985; Greeley, Lancaster, Lee, & Thomas, 1992; Sagan & Pollack, 1967). Using Equation 1, these equate to surface stress values ranging from 45–125 mN m$^{-2}$ in a 10
mb atmosphere, to 320–500 mN m$^{-2}$ in a 5 mb atmosphere. These values are high and the ranges large in part because the density of the atmosphere near the surface varies widely with relatively small altitude changes, and the threshold frictional velocity is density-dependent (Haberle et al., 2003). However, Equation 2 informs that even the lower estimates would require 1.5 m wind velocities $> 36$ m s$^{-1}$, which is greater than the maximum 1.5 m wind speeds recorded on Mars (just over $\sim 20$ m s$^{-1}$; Swann et al., 2020).

The experimentally-derived frictional velocities quoted above are problematic and yet, windblown sand and dust continue to be observed on Mars (Swann et al., 2020). In light of this discrepancy, Swann et al. (2020) recently reevaluated the threshold frictional velocity for sand-size particles on Mars and the new estimates are in better agreement with observations. Swann et al. (2020) estimate threshold frictional velocities on Mars are 0.36–0.46 m s$^{-1}$ for $\sim 100$ µm particles and 0.63–0.81 m s$^{-1}$ for $\sim 200$ µm particles (Swann et al., 2020). Using the relationship between frictional velocity and surface stress in Equation 1, these estimates give surface stress thresholds of 2.6–4.2 mN m$^{-2}$ for $\sim 100$ µm particles and 7.9–13 mN m$^{-2}$ for $\sim 200$ µm particles.

The threshold frictional velocity described here is called the “fluid” threshold because it is the threshold for mobilizing sand-size particles from rest. Recent research suggests that, on Mars, the threshold frictional velocity required to sustain saltation after it has been initiated is as much as ten times lower than the fluid threshold (Kok, 2010). This lower threshold, called the “impact” threshold, would make mobilizing sand-size particles even easier. It also lends merit to the hypothesis that saltation is the primary means of mobilizing dust particles on Mars. It has long been suspected that when saltating sand-size particles impact the surface, kinetic energy is transferred to fine dust particles that then enter into suspension (Newman et al., 2002). If the impact threshold is
significantly lower than the fluid threshold, then saltation is very likely the catalyst for
dust lifting on Mars, and this is especially likely if the recent estimates of the fluid
threshold from Swann et al. (2020) are accurate.

2.3.4 Dust source regions.

Dust lifting occurs all over Mars but there are several locations that serve as preferred
dust storm source regions. These include Acidalia, Utopia, and Arcadia in the north,
Hellas in the south, and the polar cap edges in both hemispheres (Cantor, 2007; Cantor et
al., 2001; Martin & Zurek, 1993; Wang & Richardson, 2015). MGCM simulations have
confirmed the presence of strong surface wind stresses upwind of the Acidalia and Hellas
source regions and near the seasonal CO$_2$ ice cap edges (Kahre et al., 2006; Newman et
al., 2002). As mentioned in Section 2.3.2, regional storms are especially common along
the CO$_2$ cap edges. For example, between $L_s = 107^\circ$–$274^\circ$ of MY24, there were hundreds
of local storms and seven regional-scale circumpolar storms that were observed along the
southern seasonal CO$_2$ cap edge (Cantor et al., 2001). The B storm also occurs in this
region and we are therefore especially interested in dust activity that initiates along the
southern seasonal CO$_2$ cap edge.

The retreating seasonal CO$_2$ cap edge is a favorable location for dust storm initiation
because the combined effects of several processes occurring there can produce surface
stresses large enough to lift dust. One such process is a sea breeze-like circulation
(henceforth called the cap-edge breeze) in which the thermal gradient between the CO$_2$
ice cap and the dry surface beyond it induces rising motion over the warm, dry surface
and sinking motion over the cap (Cantor et al., 2001). The cap-edge breeze is described in
Burk (1976) as a circulation that likely intensifies as the subsolar point approaches the
south pole in the days and weeks leading up to southern summer solstice (Cantor et al.,
2001; Toigo & Richardson, 2002). Observations of cap-edge dust lifting in phase with the
sun (i.e. maximizing in the afternoon hours) support this theory and suggest cap-edge lifting is highly sensitive to local temperature gradients (Toigo & Richardson, 2002).

The cap-edge breeze is 2–4 times deeper and significantly stronger on Mars than on Earth (Burk, 1976). Simulated maximum wind speeds produced by the cap-edge breeze in Burk (1976) are consistently 20 m s\(^{-1}\) regardless of the depth of the circulation. According to Burk (1976), a 250 m deep and a 60 m deep cap-edge circulation yields frictional velocities of \(\sim 0.79\) m s\(^{-1}\) and 0.92 m s\(^{-1}\), respectively, which exceed estimates from Swann et al. (2020) of the fluid threshold for sand-size particles on Mars (0.36–0.81 m s\(^{-1}\)). The cap-edge breeze may be even stronger if it is augmented by synoptic-scale winds, downslope flows, or the CO\(_2\) sublimation flow, in which case it would almost certainly be sufficient for lifting dust along the cap edge (Burk, 1976; Siili, Haberle, & Murphy, 1997).

Another process capable of lifting dust along the southern seasonal CO\(_2\) cap edge is the katabatic wind. Slope flows are common on Mars in the form of katabatic winds which are nighttime-specific downslope flows of very cold, dense air off the CO\(_2\) cap (Cantor et al., 2001). These winds are intensified in regions of steep topography and over icy surfaces regardless of the length of the slope (Savijärvi & Siili, 1993). This makes the southern CO\(_2\) cap an ideal location for the development of katabatic winds, especially in early and mid-southern summer when the elevation of the high southern latitudes becomes steeper as the seasonal CO\(_2\) cap retreats (Piqueux et al., 2006; Siili, Haberle, Murphy, & Savijärvi, 1999; Toigo & Richardson, 2002).

There are very few quantitative estimates of the magnitude of the katabatic winds at the south pole (Rafkin, Spiga, & Michaels, 2017). However, some theoretical approximations of the magnitude of the katabatic wind have been calculated using data from the Curiosity rover, located in Gale Crater at an elevation of \(-4500\) m, and Viking
Lander 2, located in an area of flat terrain at -4495 m (Rafkin et al., 2017). There are also a few modeling studies directed toward this effort (see Savijärvi & Siili, 1993; Spiga, 2011). For example, simulated katabatic flows down the 0.27° slope of Arsia Mons produce wind speeds of 17 m s\(^{-1}\) at 100 m above the surface (Savijärvi & Siili, 1993). Using Equation 2, a wind velocity of 17 m s\(^{-1}\) at 100 m in altitude equates to a surface frictional velocity of 0.5 m s\(^{-1}\) which is comparable to the fluid threshold estimated by Swann et al. (2020). However, katabatic flows can be weaker in the presence of dust and at higher latitudes which could be problematic for dust lifting near the cap edge (Savijärvi & Siili, 1993; Spiga, 2011). Nevertheless, several studies hypothesize that katabatic flows contribute to increasing surface stresses along the southern CO\(_2\) cap edge by augmenting the cap-edge breeze (see Cantor et al., 2001; Kauhanen, Siili, Järvenoja, & Savijärvi, 2008; Leovy, Zurek, & Pollack, 1973; Siili et al., 1999; Toigo & Richardson, 2002).

The CO\(_2\) sublimation flow is the third mechanism that could contribute to dust lifting along the southern seasonal CO\(_2\) cap edge during the B storm. The CO\(_2\) sublimation flow is the near-surface wind that develops when sublimating CO\(_2\) in one hemisphere and depositing CO\(_2\) in the other causes a global mass flux of CO\(_2\) toward the winter hemisphere (Toigo & Richardson, 2002). The sublimation flow produces relatively weak vertically averaged meridional winds (\(\sim 0.5\) m s\(^{-1}\)), but the Coriolis force can turn and accelerate these winds to \(\sim 10\) m s\(^{-1}\) (Barnes et al., 2017). Additionally, the CO\(_2\) sublimation flow is strengthened by the pressure gradient across the south pole, which is maximized in the southern summer season due to the combined effects of increased incident solar radiation and the amount of CO\(_2\) ice available for sublimation (Barnes et al., 2017; Toigo & Richardson, 2002). For example, at \(L_s = 255^\circ\), Toigo and Richardson (2002) show the sublimation flow has a small but noticeable effect on dust lifting around the southern seasonal CO\(_2\) cap edge at \(L_s = 255^\circ\). Though the sublimation flow alone is unlikely to produce surface winds above the fluid threshold, it has been predicted that the
CO$_2$ sublimation flow could augment the cap-edge breeze more effectively than katabatic flows (Siili et al., 1997; Toigo & Richardson, 2002).

### 2.3.5 Dust lofting.

Dust lofting describes the raising of dust that is already suspended in the atmosphere to higher altitudes. Mechanisms responsible for lofting dust play a crucial role in regulating the flux of dust and momentum between the PBL and the free atmosphere (Read et al., 2017). Some of the known mechanisms of dust lofting include the solstitial Hadley cell, rocket dust storms, and the solar escalator effect. The solstitial Hadley cell was defined and described in Section 2.2 and will not be redefined here. Instead, this section focuses specifically on the ability of these mechanisms to loft dust in the B storm.

Radiative heating of airborne dust in the rising branch of the Hadley cell intensifies upward winds and forces the Hadley cell to accelerate (Heavens et al., 2011; Wilson, 1997). Increased dust opacities during southern summer amplify this effect such that dust lofted by the Hadley cell is transported over large distances at altitudes as high as $\sim 40$ km (Haberle & Leovy, 1982). The global dispersion of dust by the Hadley cell enables GDS development (Haberle, 1986). During a GDS, the depth and intensity of the Hadley cell doubles and the descending branch shifts as much as $\sim 10–15^\circ$ poleward which enables dust to be dispersed over even greater distances (Haberle & Leovy, 1982).

The radiative-dynamic feedbacks of airborne dust in the Hadley cell are not limited to GDSs. In fact, the dynamic warming observed in the winter hemisphere opposite the A and C storms in Figure 1 illustrate that dust-induced heating occurs during regional dust storms as well (Haberle & Leovy, 1982). However, mid-level dust must be near the rising branch of the Hadley cell in order for these effects to occur. At $L_\phi = 267^\circ$ when the B storm occurs, the rising branch is $> 40^\circ$ north of the B storm (see Figure 8) and it is therefore unlikely to be lofting dust during the B storm.
The second dust lofting mechanism introduced here is rocket dust storms. These are deep convective events that occur at low latitudes primarily during the clear season (late southern summer through late southern winter). These deep convective storms are $\sim 60$ km in diameter and are capable of injecting dust to altitudes of 30–50 km (Spiga, Faure, Madeline, Määttänen, & Forgét, 2013). Once these storms perturb the atmosphere, a dense plume of dust forms that lasts several hours and produces updrafts greater than 3 m s$^{-1}$ and as high as 8–10 m s$^{-1}$ (Spiga et al., 2013). Eventually, horizontal winds force the plume into an elongated detached dust layer.

Rocket dust storms are strengthened by dust-absorbed solar radiation and can yield heating rates on the order of 15–20 K per hour (Spiga et al., 2013). This intense radiative heating results in vertical velocities that are more than twice the magnitude of the sedimentation rate of dust and therefore the upward transport of dust is maintained (Spiga et al., 2013). Without a stable stratospheric layer of air to halt development, rocket dust storms on Mars tend to continue growing until sundown (Spiga et al., 2013).

The third dust lofting mechanism described here is the solar escalator effect, which is so named for the step-like trajectory of ascending dust layers over a period of a few sols (Daerden et al., 2015). The effect occurs as a result of the diurnal cycle of localized heating of suspended dust in the atmosphere. During the day, airborne dust absorbs solar radiation and significantly warms the air in which the suspended dust resides. This causes buoyant instability which results in greater vertical mixing and rising air. At night, weak radiative cooling at longer (IR) wavelengths slows the ascent of the dust layer but does not reverse it, thus creating the ascending, stepwise trajectory of the dust layer for which the solar escalator effect is named (Daerden et al., 2015).

In a back-trajectory analysis of detached dust layers observed over the Phoenix landing site, Daerden et al. (2015) note that the constant daylight received at the high
northern latitudes during the northern summer solstice season was significant for offsetting radiative cooling at night and keeping the dust layer aloft. We suspect that the fact that the sun does not set over the high southern latitudes during the southern summer solstice season produces similarly significant effects on dust lofting. Thus, the solar escalator effect may prove especially important for dust lofting in the B storm.
3 Observations

3.1 Methods

Multi-annual observations of global atmospheric temperature, dust, and water ice on Mars have been recorded almost continuously for the last Mars decade (~ 22 Earth years; Zurek, 2017). This global climatology began in 1998 when the Thermal Emission Spectrometer (TES) on board the Mars Global Surveyor (MGS) began operations. Observations continue today with the Mars Climate Sounder (MCS) which is on board the Mars Reconnaissance Orbiter (MRO). TES documented global, vertical temperature profiles, column water vapor abundance, and column dust and water ice opacities from $L_s = 102^\circ$ in MY24 to $L_s = 82^\circ$ in MY27 (Hinson et al., 2004; Kass et al., 2016). After the decommissioning of MGS, there was a brief, MY-long gap in observations before MRO successfully entered orbit and MCS began operations. Early in MY28 ($L_s = 110^\circ$), the MCS instrument began recording global, vertical profiles of temperature, and dust and water ice extinctions (Kass et al., 2016). MCS has successfully operated for over five MY and continues taking observations today. In this section, we present an overview of the TES and MCS instruments including their architecture, observational capabilities, and uncertainties.

3.1.1 The thermal Emission Spectrometer.

TES was an instrument on board MGS, a polar orbiting satellite in a 2 AM/2 PM fixed local time orbit (Bandfield, Wolff, Smith, Schofield, & McCleese, 2013). MGS had an instantaneous field of view of 3x9 km at ~ 380 km above the surface (Bandfield et al., 2013). TES consisted of: an infrared spectrometer observing between 5.8–50 µm; a broadband thermal radiometer measuring between 5.1–150 µm; and a visible (or near-IR) radiometer measuring between 0.3–2.9 µm (Christensen et al., 2001). TES combined high spatial (~ 3 km horizontal) and spectral resolution with broad spatial (global) and spectral
coverage which enabled the instrument to distinguish between atmospheric constituents and to measure the spectral qualities of dust in the atmosphere (Christensen et al., 2001; McCleese et al., 2007).

With an orbital period of 118 minutes, MGS completed 12 rotations per day and provided north-south strips of data every $\sim 29^\circ$ in longitude (Hinson et al., 2004). The TES instrument primarily acquired nadir observations but it also performed limb scans every $\sim 3$ minutes, or about every $10^\circ$ in latitude. These were staggered by $5^\circ$ in latitude each orbit to maximize latitudinal coverage (Bandfield et al., 2013; Christensen et al., 2001; Shirley et al., 2015). Limb scans acquired temperature data at $\sim 10$ km vertical resolution from 30–65 km in altitude (Hinson et al., 2004; M. D. Smith, Pearl, Conrath, & Christensen, 2001). The temperature data used in this work were retrieved on the nadir between the surface and $\sim 11$ Pa and on the limb between $\sim 83–1$ Pa (Conrath et al., 2000). Although aerosol opacities were collected in both the nadir and limb geometries, only those data retrieved on the nadir view are used in this work (Conrath et al., 2000).

The temperature and dust retrieval algorithms for TES are detailed in Conrath et al. (2000) and M. D. Smith, Pearl, Conrath, and Christensen (2000), respectively. Here, we focus our discussion on the sources of error in the TES temperature and dust retrievals. The uncertainties in the temperature data are largest near the surface where both ground and atmospheric spectral radiances are observed by TES. Near-surface uncertainties also arise because the 10 km vertical resolution is too coarse to resolve the small vertical temperature variations that are common near the surface (Hinson et al., 2004). Finally, assumptions made in the surface pressure calculation can introduce errors in the temperatures retrieved in the lowest $\sim 10$ km of the atmosphere, especially in warmer environments (Conrath et al., 2000). These errors can affect temperatures by as much as
∼ 6 K near the surface but become insignificant above 10 km (Conrath et al., 2000; Hinson et al., 2004).

We do not anticipate near-surface errors in the TES data to significantly affect our results because the thermal effects of the B storm are located between 200–20 Pa. We expect the most significant source of error in our data comes from the estimated CO₂ absorption coefficients used in the TES retrieval algorithm (Hinson et al., 2004). The magnitude of this uncertainty is < 2 K for the bulk of the atmosphere above one scale height (Conrath et al., 2000; Hinson et al., 2004; M. D. Smith et al., 2001). We therefore expect our analysis of the mid-level temperature structure as observed by TES to be accurate to within ∼ 2 K.

Aerosol opacities retrieved by TES were limited by a surface temperature threshold that allowed retrievals to be made only in areas where surface temperatures exceeded 220 K (M. D. Smith et al., 2001). This ensured that the thermal contrast between the ground and the near-surface environment was large enough for accurate retrievals to be made. Aerosol opacities were retrieved as column total abundances, and later separated into dust and water ice components using a least squares fit. Uncertainties in these retrievals are typically on the order of 0.05 for opacities < 0.5 (M. D. Smith et al., 2001). This margin grows with increasing optical thickness because the retrieval algorithm assumes aerosols are non-scattering and well-mixed with CO₂ in the atmosphere.

The assumptions in the retrieval algorithm are fair for background dust levels but cause greater uncertainties in the opacities retrieved during dust storms. As is explained in M. D. Smith et al. (2001), dust storm opacities retrieved by TES are therefore better understood in terms of relative optical thickness rather than absolute measures of the amount of suspended dust in the atmosphere. M. D. Smith (2004) later improved the uncertainty estimate such that TES-retrieved aerosol optical depths are now estimated to
be accurate to ±0.05 or 10% of the total optical depth, whichever is greater. Averaging the data can reduce this uncertainty by a factor of two (M. D. Smith, 2004). Using these estimates, we calculate that our zonal mean 50 Pa column dust opacities from TES are accurate to < 0.03.

3.1.2 Mars Climate Sounder.

MCS is an instrument on board MRO, a sun-synchronous, polar orbiting satellite following a 3 AM/3 PM local time orbit (Bandfield et al., 2013). MRO has an instantaneous field of view of 1.5x8 km at a mean altitude of ~280 km (Bandfield et al., 2013). MRO completes one orbit every 112 minutes and 12 seconds and provides north-south strips of data every ~27° in longitude (McCleese et al., 2007). Unlike TES, MCS observes primarily on the limb. Limb scans are performed every 34 seconds or every 1.86° in latitude (~110 km). MCS measurements on the limb provide profiles of temperature and aerosol abundance at high resolution. MCS retrievals have a 10 km vertical resolution in the lowest scale height of the atmosphere, and a 5 km vertical resolution at altitudes between 10–80 km (McCleese et al., 2007).

MCS is an IR radiometer with nine spectral channels that each have a 21-element linear detector array (Shirley et al., 2015). This design enables high vertical resolution profiles of temperature, dust extinction rate, and condensate opacity to be taken by MCS (Shirley et al., 2015; McCleese et al., 2007). Retrievals are made using the limb-staring method, which is a technique that reduces data gaps and the signal-to-noise ratio, both of which are common in data retrieved by limb scanning methods (McCleese et al., 2007). At the poles, MCS performs more frequent nadir and off-nadir observations, which provide a more comprehensive picture of the diurnal and seasonal temperature variations at the high latitudes on Mars (McCleese et al., 2007).
MCS data are somewhat limited in areas where the vertical temperature gradient is strong, such as over the winter pole (Shirley et al., 2015). However, systematic errors in the MCS retrievals at the winter pole have improved since a 2D radiative transfer scheme was implemented into the retrieval algorithm in order to correct for horizontal gradients in the IR radiances retrieved on the limb (Kleinbühl, Friedson, & Schofield, 2017; Kleinbühl et al., 2009).

The path length of the off-nadir view often limits the frequency of MCS retrievals when aerosol extinction rates are $> 4 \times 10^{-3}$ km, such as during major dust events (Shirley et al., 2015). This is potentially more problematic over the polar regions where off-nadir retrievals are performed more frequently. The MCS profiling sequence is designed so that limb observations can be combined with nadir observations that were taken earlier in the orbit, which improves the retrieval accuracy of the MCS instrument. Although a mechanical issue early on in the mission prevented nadir observations from being performed for several months in MY28, the instrument failure does not affect our investigation because MY28 is a GDS year and is therefore excluded from our study (Shirley et al., 2015).

MCS temperature uncertainties generally range from $\sim 0.5$–2 K overall (Kleinbühl et al., 2009; Shirley et al., 2015). The lower estimate is applicable for observations made over the northern hemisphere midlatitudes during northern summer. The higher estimate is applicable to observations made below 10 km in regions where the atmosphere is highly opaque, and to observations made above 60 km where the signal-to-noise ratio is reduced (Kleinbühl et al., 2009). In the southern winter season, uncertainties are on the order of $\sim 0.5$ K between 10–40 km, increase to $> 1$ K between 40–65 km, and maximize at $> 3$ K above 65 km (Kleinbühl et al., 2009). Our study concerns temperatures at $\sim 50$
Pa (\( \sim 25 \text{ km} \)) at the south pole in the peak summer season, and we therefore estimate the errors in our MCS-retrieved temperatures are \(< 2 \text{ K} \).

The MCS retrieval algorithm calculates the dust extinction rate \( (d_z \tau, \text{ km}^{-1}) \) at a path length of 1 km using Mie theory and the refractive indices of dust particles measured by TES on MGS and miniTES on the Mars Exploration Rovers (for a detailed description of the MCS retrieval algorithm, see Kleinböhl, Schofield, Abdou, Irwin, & de Kok, 2011; Kleinböhl et al., 2009). For the purposes of this study, extinction rate is converted to mass mixing ratio \( (q, \text{ ppm}) \) to enable the direct comparison of MCS data to MGCM output. The conversion is given by Heavens et al. (2011):

\[
q = \frac{4}{3} \frac{\rho_D}{Q_{\text{ext}}} d_z \tau \rho_{\text{atm}} r_{\text{eff}}
\]  

(3)

where the mass of the dust particles in a given volume, \( \rho_D \), is \( 3 \times 10^3 \text{ kg m}^{-2} \) and \( \rho_{\text{atm}} \) is the density of the atmosphere at altitude. The MCS retrieval algorithm assigns a dust extinction efficiency, \( Q_{\text{ext}} \), of 0.35, and an effective particle radius, \( r_{\text{eff}} \), of 1.06 \( \mu \text{m} \) (Kleinböhl et al., 2009). Using these values, Equation 3 can be simplified to

\[
q = 1.2 \times 10^4 \frac{d_z \tau}{\rho_{\text{atm}}}
\]  

(4)

Converting dust extinction rate to mass mixing ratio is especially useful for diagnosing where dust is accumulating in the atmosphere (Heavens et al., 2011). Dust extinction rates represent the amount of radiation absorbed and scattered by dust particles in the atmosphere, and while this is useful for understanding the radiative properties of airborne dust, it provides little information on the amount and location of that airborne dust. Mass mixing ratio, on the other hand, represents the mass of dust present in a kilogram of air. This is useful for identifying the altitude to which dust is lofted in the B storm. The conversion between extinction rate and mass mixing ratio can introduce
additional errors to the data due to uncertainties in the particle size distribution estimated for the Martian atmosphere. However, for a reasonable size range of particles ($0.75 \mu m \leq r_{\text{eff}} \leq 6.00 \mu m$), the ratio of $Q_{\text{ext}}$ to $r_{\text{eff}}$ varies by $<30\%$, and thus such errors are small (Heavens et al., 2011).

MCS dust retrievals are complicated by an incomplete understanding of how dust particle size, composition, and scattering factor into dust spectroscopy (Shirley et al., 2015). Although these errors are difficult to quantify, dust extinction rates are typically considered accurate to $10^{-5}$ km$^{-1}$ (Kleinböhl et al., 2009). For extinction rates $>10^{-5}$ km$^{-1}$, uncertainties are no greater than $10\%$ (Heavens et al., 2011). These errors cannot be converted to mass mixing ratio because the atmospheric density parameter, $\rho_{\text{atm}}$, in Equation 4 varies depending on the temperature and pressure at altitude. Additionally, the uncertainty for extinction rates exceeding $10^{-5}$ km$^{-1}$ is dependent on the mass of dust in the atmosphere. To account for these effects, the error estimates from Kleinböhl et al. (2009) and Heavens et al. (2011) were applied to the MCS extinction rates before they were converted to mass mixing ratios. We performed this analysis on the MCS data retrieved south of 20° S between $L_s = 247°–287°$ in MY29–MY33, and found the average error for dust concentrations retrieved in the lowest two scale heights of the atmosphere is $\sim 0.553$ ppm $\pm 0.293$. However, this uncertainty is greatest in dusty regions such as in the southern midlatitudes, which are tens of degrees in latitude north of the observed B storm.

### 3.1.3 Instrument comparison: MCS and TES.

MCS and TES temperature retrievals are comparable, generally differing by $<3$ K after accounting for local time differences (Bandfield et al., 2013). Most of the discrepancies between the retrievals appear at the top and bottom of the observed atmosphere. For example, TES retrievals above 1 Pa are consistently $>10$ K colder than MCS retrievals (Shirley et al., 2015). Near the surface (below 610 Pa or $\sim 3$ km), TES
retrievals are consistently ∼5 K warmer than MCS retrievals (Shirley et al., 2015). However, Shirley et al. (2015) found no systematic biases in the temperatures retrieved between 610–1 Pa, and in fact assert that temperatures are in especially good agreement below 20 Pa. Our study is focused on temperature observations at 50 Pa (∼25 km) and our assessment of the thermal structure of the B storm is limited to the column between 610–15 Pa, which is within the region where TES and MCS retrievals are in good agreement. We are confident that the TES and MCS temperatures presented here agree to within ∼3 K of one another and that the primary source of error between them is the local time difference (∼1 hour).

3.2 Results

The global climatology from MCS and TES includes atmospheric temperature and dust observations from MY24 to the present day. However, data retrieved during four of those years are excluded from this study: MY25, MY27, MY28, and MY34. Three of those years, MY25, MY28, and MY34, are GDS years that must be excluded from this work because the A, B, and C regional storms are not observed during GDS years. MY27 is excluded because there is little available data from that year between the decommissioning of MGS and the successful insertion of MRO. As a result, this study uses data retrieved in the following seven years in order to characterize the observed structural evolution of the B storm: MY24 and MY26 from TES, and MY29–MY33 from MCS (see Table 1).

The TES data were provided by Conrath et al. (2000) and M. D. Smith (2004). Dust optical depth and atmospheric temperature were retrieved on the nadir between ∼610–11 Pa, and these data are binned by 5° in L\textsubscript{s}, 3° in latitude, and 7.5° in longitude. The MCS data are from the derived data product available on NASA’s Planetary Data System (PDS)
and include dust extinction rates and atmospheric temperatures between $\sim 610–1$ Pa. These data are binned by $10^\circ$ in latitude, $10^\circ$ in longitude, and $10^\circ$ in $L_s$.

### 3.2.1 Overview of B storm observations.

We begin our observational analysis of the B storm by reproducing Figure 2 from Kass et al. (2016), which shows 50 Pa temperatures for MY24, MY26, and MY29–MY32, and adding the MY33 B storm to the analysis. Our results are presented in Figure 9. The zonal mean 50 Pa temperatures retrieved during the dusty season in MY24, MY26, and MY29–MY32 are in good agreement with those from (Kass et al., 2016) in Figure 2. The A, B, and C storms present at the same time as in Figure 2, and the weak C storms in MY24 and MY30 appear in our data as well.

We extended the analysis from Kass et al. (2016) through MY33 and found the regional storms in MY33 have similar characteristics to those observed in earlier years. Figure 9 confirms the MY33 A, B, and C storms are associated with temperatures $> 200$ K which is the threshold temperature defining the regional storms in Kass et al. (2016). In all seven MYs, the A and C storms develop at $\sim 45^\circ$ S in the southern spring season ($L_s = 215^\circ–250^\circ$) and the late southern fall season ($L_s = 315^\circ–340^\circ$), respectively. The A and C storms produce a northern hemisphere dynamical response at $55^\circ$ N that raises 50 Pa temperatures upwards of 200 K. In all seven MYs, the B storm occurs in the highest southern latitudes during southern summer solstice ($L_s = 270^\circ$), and does not cause dynamic warming in the northern hemisphere. However, the dynamical effects of the A storm linger well into the onset of the B storm, especially in MY29 and MY30. The warmest 50 Pa B storm temperatures are located over the receding CO$_2$ cap which is represented by a dashed white line in Figure 9 according to the observations presented in Piqueux et al. (2006).
Figure 9. As in Figure 2 but includes MY33.

The A, B, and C regional storms are well represented in the zonal mean dust fields shown in Figures 10 and 11. Zonal mean 50 Pa dust mixing ratios for MY29–MY33 are shown in Figure 10 and zonal mean column dust opacities for MY24 and MY26 are shown in Figure 11. In Figure 10, the A and C storms produce dust mixing ratios > 12 ppm and the B storm produces dust mixing ratios as high as ∼ 8 ppm at 50 Pa. The highest B storm dust mixing ratios are > 8 ppm and they occur in MY31 and MY33. Column dust opacities in Figure 11 exceed 0.4 during all three regional storms and the highest column dust opacities produced during the B storm are located south of the
receding CO$_2$ cap edge. Figure 10 also shows that B storm dust concentrates over the receding CO$_2$ cap edge at 50 Pa. As in Figure 9, the CO$_2$ cap edge is denoted by the dashed black line in Figures 10 and 11.

Figure 10. MCS-observed zonal mean daytime dust mixing ratios at 50 Pa over $L_s = 180^\circ$–$360^\circ$ in MY29–MY33. Contours are labeled inline. The dashed line indicates the seasonal CO$_2$ cap edge as observed by Piqueux et al. (2006).
Figure 11. As in Figure 10 but for TES column dust opacities retrieved in MY24 and MY26.

3.2.2 Observed phases of the B storm.

To better understand the evolution of the B storm, we analyze the dayside and nightside 50 Pa temperatures and dust mixing ratios in 10° $L_s$ increments over the duration of the B storm ($L_s = 247°–287°$). The term “dayside” (“nightside”) refers to the 2 PM (2 AM) local time observations from TES and the 3 PM (3 AM) local time observations from MCS, but the sun is always up at the high southern latitudes around southern summer solstice ($L_s = 270°$). Dayside and nightside 50 Pa temperatures are shown for all non-GDS years in Figures 12 and 13. The dayside and nightside 50 Pa dust mixing ratios are shown in Figures 14 and 15, but are only available for years in which MCS performed observations (MY29–MY33). The data in Figures 12–15 are binned by 10° of $L_s$ and shown at 10° of $L_s$ intervals: $L_s = 247°$, 257°, 267°, 277°, and 287°. In the following paragraphs, we present a thorough description of the observed structure of the B storm as it evolves over time and we define the growth ($L_s = 247°–257°$), peak ($L_s = 267°$), and decay ($L_s = 277°–287°$) phases of the observed B storm.
Figure 12. Observed dayside south polar 50 Pa temperatures at (columns) $L_s = 247^\circ$, 257°, 267°, 277°, and 287° for (rows) MY24 and MY26 (2 PM) and MY29–MY33 (3PM). Data are binned in 10° of $L_s$ increments. Dashed line represents the seasonal CO$_2$ cap edge from Piqueux et al. (2006).
Figure 13. As in Figure 12 but for data retrieved on the nightside (TES=2 AM, MCS=3 AM).

The growth phase of the B storm is captured in the first two columns \((L_\parallel = 247^\circ\) and \(L_\parallel = 257^\circ\)) of Figures 12–15. At \(L_\parallel = 247^\circ\), the B storm is in its earliest stages of
development in all but MY29 and MY30. Dayside 50 Pa temperatures are $< 200 \text{ K}$ over the pole and $> 200 \text{ K}$ north of the seasonal CO$_2$ cap. In MY29 and MY30, dayside temperatures are warm ($> 210 \text{ K}$) across the polar region as the late season A storms dissipate. Meanwhile, dayside temperatures are $> 200 \text{ K}$ throughout the rest of the southern hemisphere. The A storm can be seen merging into the B storm in the MY29 and MY30 50 Pa temperature retrievals shown in Figure 9. The nightside temperatures shown in Figure 13 indicate that the southern high latitudes remain warmer than 200 K throughout the night in MY29 and MY30. In MY26, MY32, and MY33, however, nightside 50 Pa temperatures are $\sim 190–195 \text{ K}$ across the southern hemisphere.

The 50 Pa dust fields at $L_s = 247^\circ$ in Figures 14 and 15 show the onset of the B storm. The dayside 50 Pa dust mixing ratios in MY31 and MY33 show dust is present poleward of 60$^\circ$ S. This dust appears detached from the midlatitude dust also present at this time. In MY31 and MY33, dust concentrations are $> 4 \text{ ppm}$ poleward of 60$^\circ$ S in the eastern hemisphere between 45$^\circ$ E and 135$^\circ$ W. On the nightside (Figure 15), dust concentrations increase to $> 6 \text{ ppm}$. There is limited evidence of B storm development in MY29, MY30, and MY32 because the lingering A storm somewhat masks the development of the B storm in MY29 and MY30 and the B storm begins about 5$^\circ$ of $L_s$ later than usual in MY32.

By $L_s = 257^\circ$, the MY24, MY26, MY30, MY31, and MY33 B storms are well into their growth phases, the MY30 and MY32 B storms are beginning to develop, and the MY29 B storm continues to be delayed by the late season A storm. Figures 12 and 13 show the lingering effects of the MY29 A storm are still present at $L_s = 257^\circ$ but have weakened significantly. Dayside 50 Pa temperatures in Figure 12 indicate warm ($> 200 \text{ K}$) mid-level air throughout the southern hemisphere in all seven MYs, and nightside 50 Pa temperatures in Figure 13 show warm ($> 200 \text{ K}$) air in the polar region south of 60$^\circ$ S.
Only the region over the residual cap is < 200 K on the nightside in MY24, MY29, and MY32.

Figure 14. Observed dayside 50 Pa dust mixing ratios over the south pole at (columns) $L_s = 247^\circ$, 257°, 267°, 277°, and 287° for (rows) MY29–MY33 (MCS=3PM). The data are binned in 10° of $L_s$ increments. The dashed line indicates the seasonal CO$_2$ cap edge from Piqueux et al. (2006).

On the dayside at $L_s = 257^\circ$, 50 Pa dust concentrations are $> 4$ ppm poleward of 60° S in MY29–MY33, but the MY29 dust appears to be A storm dust lingering in the 60°–70° S latitude band (Figure 14). On the nightside, dust concentrations are higher, around $\sim 6$ ppm, over parts of the polar region in all seven MYs. MY31 is especially
dusty in the eastern hemisphere between 45° E and 135° W where dust mixing ratios exceed \( \sim 10 \) ppm on the dayside and \( \sim 12 \) ppm on the nightside. MY30, MY32, and MY33 are also dustier in this region but maximum dust mixing ratios in these years are significantly lower (\( \sim 5 \) ppm) than in MY31.

Figure 15. As in Figure 14 but for nightside retrievals (3 AM).

The B storm peaks in intensity just before southern summer solstice (\( L_s = 270° \)) at around \( L_s = 267° \). Warm temperatures and high dust mixing ratios are more uniform around the pole during peak intensity, and dayside and nightside 50 Pa temperatures are warmest directly over the south pole in MY24, MY26, and MY30–MY33. In MY29,
dayside temperatures in Figure 12 are still fairly uniform throughout the southern hemisphere, but nightside temperatures in Figure 13 are more indicative of B storm temperatures in the other six years.

Dust mixing ratios at 50 Pa are maximized on the nightside at $L_s = 267^\circ$ (Figure 15). While dayside dust concentrates south of $\sim 80^\circ$ S in MY29, 70° S in MY30, MY32, and MY33, and 60° S in MY31, nightside dust concentrates 10° further north at 70° S in MY29, 60° S in MY30, MY32, and MY33, and nearly 50° S in MY31. Additionally, 50 Pa dust concentrates in the eastern hemisphere between 45° E and 135° W at $L_s = 267^\circ$ on the dayside in all seven MYs, but especially in MY29, MY30, and MY33 (Figure 14). The 50 Pa dust field is somewhat more evenly distributed around the pole on the nightside than on the dayside, but the concentration of dust in the eastern hemisphere is present at both times of day.

The dissipation of the B storm begins immediately following peak intensity and continues through $L_s = 277^\circ$–287°. In Figure 12, dayside temperatures north of 70° S are $\sim 10$ K cooler at $L_s = 277^\circ$ than at $L_s = 267^\circ$, and in MY24, MY26, and MY30–MY32 the northernmost extent of the warmest dayside temperatures ($\sim 215$ K) decrease by more than 10 K between $L_s = 267$–277°. On the nightside in Figure 13, 50 Pa temperatures are $> 200$ K poleward of 60° S at $L_s = 277^\circ$, which is 5°–10° further south than at peak intensity. The dayside 50 Pa dust field indicates that dust mixing ratios at $L_s = 277^\circ$ are about half the magnitude they were at $L_s = 267^\circ$. The northernmost extent of the dust continues to be located $\sim 10^\circ$ further north on the nightside than on the dayside, however, dust concentrations are no higher on the nightside than on the dayside after peak intensity.

The B storm cessation phase continues through $L_s = 287^\circ$. By this time, most of the dust that was present at 50 Pa at $L_s = 267^\circ$ has fallen out of suspension. Dayside and nightside dust mixing ratios are only $\sim 1$–2 ppm larger at the pole than in the
midlatitudes at $L_s = 287^\circ$ whereas dust mixing ratios had been as much as $\sim 6$ ppm larger at the pole than in the midlatitudes at peak intensity ($L_s = 267^\circ$). Dayside and nightside 50 Pa temperatures are still $> 200$ K over the pole at $L_s = 287^\circ$, but the latitudinal extent of the warm temperatures ($> 200$ K) is now similar on the nightside and the dayside. However, the meridional temperature gradient is larger on the nightside than on the dayside. In MY24, for example, 50 Pa temperatures are as much as $\sim 10$ K warmer over the pole and $\sim 10$ cooler north of $60^\circ$ S on the nightside than on the dayside.

### 3.2.3 Vertical profiles of temperature and dust.

TES- and MCS-retrieved vertical profiles of temperature are shown in Figure 16. TES did not retrieve vertical profiles of dust so the dust mixing ratio cross-sections in Figure 17 are from MCS-observing years alone (MY29–MY33). Each plot is centered over the south pole and shows nightside retrievals on the left half of the plot and dayside retrievals on the right. This mimics the orbit of the MGS and MRO spacecraft which descend on the nightside, cross over the south pole, then ascend on the dayside during each orbit.
Figure 16. Observed zonal mean temperature cross-sections over the south pole at (columns) $L_s = 247^\circ$, $257^\circ$, $267^\circ$, $277^\circ$, and $287^\circ$ degree for (rows) MY24, MY26, MY29–MY33. The data are binned in 10° $L_s$ increments. The left side of the plot shows the nightside (3 AM) retrievals and the right side shows the dayside (3 PM) retrievals. The horizontal dotted line indicates 50 Pa.

As in Figures 12–15, the data in Figures 16 and 17 are binned by 10° of $L_s$ and are shown every 10° of $L_s$ between $L_s = 247^\circ$–$287^\circ$. Throughout this section, “warm” air
refers to temperatures ≥ 200 K and “dusty” air refers to mixing ratios ≥ 4 ppm since 50 Pa temperatures exceed 200 K and 50 Pa dust mixing ratios exceed 4 ppm during the B storm in all seven MYs.

The growth phase of the B storm is shown in the first two columns (Ls = 247° and Ls = 257°) of Figures 16 and 17. The MY29 B storm is overshadowed by the lingering A storm throughout the growth phase so the vertical distribution of dust at Ls = 247°–257° in MY29 is atypical. However, there is some evidence of a B storm-like dust column developing over ~ 70° S, and the thermal structure of the MY29 B storm resembles the thermal structure of the B storm in the other six MYs despite the lingering effects of the MY29 A storm.

In MY30–MY33, Figure 17 shows that the B storm presents as a column of dust that travels poleward from 70° S to 80° S between Ls = 247°–257°. At Ls = 247° in MY30 and MY31, the dusty (> 4 ppm) column extends to just above 50 Pa. In MY33, the dusty column extends to just below 50 Pa and, in MY32, the B storm is shallow and the dusty column is below 100 Pa. The thermal structure of the B storm at Ls = 247° presents as a column of warm (> 200 K) air that extends beyond 50 Pa over ~ 50° S. Temperatures at 50 Pa remain > 200 K overnight in MY29 and MY30, and dayside temperatures are warmer than nightside temperatures in all seven MYs. By Ls = 257°, the dusty column extends to 50 Pa in all but MY29. The depth of the warm column of air is largely unchanged at Ls = 257°, but it has migrated south to ~ 70° S in MY29 and to ~ 90° S in MY24, MY26, and MY30–MY33.

When the B storm peaks in intensity at Ls = 267°, the column of dusty warm air is centered directly over the south pole and dust mixing ratios are maximized at 50 Pa. In Figure 17, peak 50 Pa dust mixing ratios are > 4 ppm in MY29, > 6 ppm in MY30–MY32, and ~ 8 ppm in MY33. The dusty column is deep, extending to ~ 25 Pa

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in all but MY29, and it is about \( \sim 10^\circ \) wide in latitude on both the dayside and the nightside. The highest dust concentrations (> 8 ppm) are located between \( \sim 200–100 \) Pa at \( L_s = 267^\circ \), and temperatures maximize between \( \sim 215–220 \) K at 50 Pa. The warm (> 200 K) column extends to \( \sim 20 \) Pa in MY24, MY26, and MY31, \( \sim 25 \) Pa in MY30, MY32, and MY33, and \( \sim 30 \) Pa in MY29.

**Figure 17.** As in Figure 16 but for dust mixing ratios retrieved by MCS. Excludes MY24 and MY26 in which TES was the observing instrument.

During the decay phase (\( L_s = 277^\circ–287^\circ \)), dust falls out of suspension over the pole and temperatures throughout the column cool slowly. Between \( L_s = 267^\circ–277^\circ \), the B storm becomes shallower, cooler, and clearer. In MY31–MY33, the 200 K isotherm and 4
ppm isoline indicate the B storm is $\sim 10$ Pa shallower at $L_s = 277^\circ$ than at $L_s = 267^\circ$. In MY24, the warm column is relatively unchanged at $L_s = 277^\circ$, but, in MY26, the column is nearly $\sim 20$ Pa shallower at $L_s = 277^\circ$. In MY29, the B storm is actually $< 5$ Pa deeper at $L_s = 277^\circ$ than at $L_s = 267^\circ$, but the warm column becomes another $\sim 10$ Pa shallower in all seven MYs by $L_s = 287^\circ$. At the same time, the dust is $\sim 15–20$ Pa shallower in MY29, MY30, and MY32, and $\sim 50–60$ Pa shallower in MY31 and MY33.

### 3.2.4 Interannual variability.

The B storm occurs around southern summer solstice ($L_s = 270^\circ$) in all non-GDS years observed by TES and MCS, and the B storm consistently peaks in intensity at around $L_s = 267^\circ$. This is best illustrated in Figure 17 which shows the B storm is deepest, widest, and dustiest at $L_s = 267^\circ$ in MY30–MY33, and that only in MY29 is the timing of peak intensity somewhat arbitrary. The cessation of the B storm is similarly consistent, concluding by $L_s = 290^\circ$ in all seven MYs. The onset the B storm is more variable but still fairly repeatable from year to year. Zonal mean 50 Pa temperatures $> 200$ K in Figure 9 indicate the B storm occurs from $L_s = 250^\circ–290^\circ$ in MY26 and in MY31–MY33. In MY24, MY29, and MY30, B storm onset is $5^\circ$ of $L_s$ later and, in MY31, B storm onset is $5^\circ$ of $L_s$ earlier. The zonal mean dust fields also reflect the repeatability of the B storm. Zonal mean 50 Pa dust mixing ratios $> 4$ ppm in Figure 10 indicate the MY30–MY33 B storms initiate around $L_s = 250^\circ$ and decay around $L_s = 285^\circ \pm 3^\circ$ of $L_s$. In MY29, the B storm initiates after $L_s = 260^\circ$ but still decays around $L_s = 285^\circ$.

Another repeatable characteristic of the B storm is that 50 Pa temperatures and dust mixing ratios tend to maximize in the eastern hemisphere between $45^\circ$ E and $135^\circ$ W. This is most evident in MY30–MY33 at $L_s = 257^\circ$, which are shown in Figures 14 and 15. Additionally, 50 Pa dust concentrations maximize over the receding CO$_2$ cap edge at peak intensity ($L_s = 267^\circ$), as can be seen in Figure 10 between $70–80^\circ$ S. Here, zonal
mean 50 Pa dust mixing ratios are maximized along the northernmost extent of the receding CO$_2$ cap edge (indicated by the black dashed line).

The B storm shows the most interannual variability during the growth phase. Specifically, the vertical distribution of dust between L$_s$ = 247°–257° is highly variable from year to year. Figure 17 shows that the MY30 B storm has a well defined column of dust over 70° S, the MY31 and MY33 B storms are less dusty and also centered over 70° S, and the MY32 B storm is a shallow dust column that is just beginning to form. At L$_s$ = 257°, the MY30 B storm has dust mixing ratios of around ∼ 4 ppm at 50 Pa, ∼ 8 ppm at 100 Pa, and > 8 ppm at 200 Pa. While the MY32 and MY33 B storms have a similar dust profile, the columns are ∼ 1 ppm dustier in MY33 and ∼ 1 ppm less dusty in MY32. The MY31 B storm is very dusty at L$_s$ = 257° with peak mixing ratios maximizing at over 8 ppm between 200–50 Pa. However, the MY29 B storm shows the opposite pattern, becoming less dusty over the polar region between L$_s$ = 247° and L$_s$ = 257°.

3.3 Summary of Observations

Our observational analysis of the B storm confirms and builds upon the findings of Kass et al. (2016). The B storm occurs over the high southern latitudes around southern summer solstice in all non-GDS years observed by TES (MY24, MY26) and MCS (MY28–MY33). The B storm does not produce dynamic warming in the northern hemisphere, but the dynamical effects of the A storm often linger into the development of the B storm. In every MY studied here, 50 Pa temperatures are > 200 K, 50 Pa dust concentrations are > 4 ppm, and column dust opacities are > 0.4 during the B storm. Additionally, peak 50 Pa temperatures (∼ 215 K) and dust concentrations (∼ 8 ppm) are located over the receding seasonal CO$_2$ cap in all of the non-GDS years studied here.
The B storm shows little interannual variability overall. The storm typically occurs from \( L_s = 250° \pm 290° \), with time of onset varying by \(< 5° \) of \( L_s \) even when preceded by a late season A storm, such as in MY29 and MY30. The B storm peaks in intensity just before southern summer solstice (\( L_s = 270° \)) around \( L_s = 267° \) and decays by \( L_s = 290° \) regardless of when the storm initiated. In MY29 and MY30, the late season A storm weakens the B storm such that peak 50 Pa temperatures are \( \sim 5 \) K cooler and peak 50 Pa dust mixing ratios are \( \sim 2 \) ppm lower at \( L_s = 267° \) in the MY29 and MY30 B storms than in the other five B storms. The MY24 A storm initiates late (\( \sim L_s = 230° \)) but does not interfere with the development of the MY24 B storm.

In this section, three phases of the B storm were identified and defined using the TES and MCS retrievals. These are the growth (\( L_s = 247° \pm 257° \)), peak (\( \sim L_s = 267° \)), and decay (\( L_s = 277° \pm 287° \)) phases. In brief, these phases have the following characteristics. During the growth phase, a warm (\( > 200 \) K) and dusty (\( > 4 \) ppm) column of air deepens as it migrates from the southern midlatitudes toward the south pole. During the growth and peak phases, B storm temperatures are warmest on the dayside while the dust field is dustiest on the nightside, and dust tends to concentrate in the eastern hemisphere. At peak intensity, 50 Pa temperatures maximize at \( \geq 215 \) K, 50 Pa dust mixing ratios maximize at \( \sim 8 \) ppm, and dust concentrates off-center of the pole toward the eastern hemisphere. During the decay phase, around half the dust in the column falls out of suspension and temperatures cool by \( \sim 10 \) K every \( 10° \) of \( L_s \). The B storm cessation phase consistently ends by \( L_s = 290° \).

In the following section, an analysis of a B storm simulated by the high resolution MGCM is presented. The modeling study provides additional insight into the physics of the B storm that are not captured by TES and MCS. Most notably, these include the model-predicted winds and particle size distributions during the B storm. From the
simulated winds and temperatures, we can determine how the B storm interacts with the larger global circulation that develops during the southern summer season. We are also able to determine what mechanisms of dust lifting and lofting may be involved in the B storm. Additionally, the MGCM enables frequent analysis of the simulated B storm.

Whereas orbital observations of the south pole are available every $\sim 12$ hours, model data over the south pole can be available as often as every $\sim 7.7$ minutes at $\sim 1^\circ \times 1^\circ$ resolution. In our baseline simulation, data are output hourly which is necessary for resolving diurnal and semi-diurnal circulations at the south pole such as the cap-edge breeze and katabatic flows.
4 Modeling

4.1 Model Description

We use the NASA Ames MGCM to simulate the B storm at high resolution (1° x 1°; ~ 60x60 km horizontal resolution at all locations). The vertical grid is defined by a hybrid sigma-pressure coordinate system that is terrain-following near the lower boundary and isobaric aloft. The MGCM is a hydrostatic model with a cubed-sphere grid and a finite-volume (FV3) dynamical core that was adapted for Mars from the NOAA/GFDL Earth model (Bertrand et al., 2020). The cubed-sphere grid, which was introduced in Section 1.3 and illustrated in Figure 4, is highly uniform and allows the model to run at high resolution on parallel computers (Bertrand et al., 2020). The grid can be stretched so that one of the six tiles can resolve a smaller area at higher resolution, and it supports nesting on both the native and stretched tiles (Geophysical Fluid Dynamics Lab [GFDL], n.d.). The FV3 dynamical core solves the equations of motion that govern large-scale atmospheric dynamics. These are the continuity, vector momentum, and thermodynamic equations that represent the laws of conservation of mass, momentum, and energy, respectively. A hydrostatic approximation is applied to the vertical momentum equation to eliminate vertically-propagating sound waves from the numerical solutions.

Land-surface interactions, dust lifting mechanisms, and cloud microphysics are all examples of physical processes that are represented by parameterizations in the model. The physics schemes employed by the MGCM are adapted from the NASA Ames Legacy MGCM and described in detail in Haberle et al. (2019). In this section, we offer a review of the MGCM physics parameterizations relevant to this work. These include the schemes representing the surface properties of Mars (topography, thermal inertia, albedo, etc.), the PBL scheme, the radiative transfer scheme, and dust lifting scheme used in our simulation.
4.1.1 Model surface properties.

The MGCM is initialized with fixed surface topography, albedo, and thermal inertia maps. The topographical map is based on the 1/16° resolution topography retrieved by the Mars Orbiter Laser Altimeter on board MGS, and the surface thermal inertia and surface albedo maps are derived from TES observations (D. E. Smith et al., 1999; Wilson, Neumann, & Smith, 2007). The south polar residual CO$_2$ ice cap is not explicitly simulated in the MGCM, but is represented by regions of high surface thermal inertia and albedo.

The seasonal growth and retreat of the simulated CO$_2$ ice caps at the poles are self-consistently determined through a surface energy balance, which is affected by, among other parameters, surface thermal inertia, albedo, topography, and the albedo prescribed to the seasonal CO$_2$ ice caps. The nominal albedos prescribed to the simulated northern and southern seasonal CO$_2$ ice caps are 0.7 and 0.51, respectively, and these can be changed at model initialization if necessary. The nominal CO$_2$ ice albedos are chosen to produce a good fit to the timing and magnitude of the seasonal exchange of CO$_2$ between the surface and the atmosphere each season, which is quantified by comparing the observed (by the Viking landers) and simulated annual surface pressure cycles.

4.1.2 The planetary boundary layer.

The MGCM planetary boundary layer (PBL) scheme predicts wind and temperature profiles in the lowest scale height of the atmosphere and mixes tracers throughout the boundary layer (Bertrand et al., 2020). The PBL scheme does not account for the effects of non-local mixing in the lower atmosphere and such deficiencies are likely responsible for weak vertical transport within the simulated PBL (see Haberle et al., 2019). Surface heat and momentum fluxes are calculated in the PBL scheme according to Monin-Obukhov similarity theory (Kahre, Haberle, Hollingsworth, & Wolff, 2020).
Near-surface heat fluxes are handled by the level-2 Mellor and Yamada (1982) turbulence closure scheme, which has been adapted for Mars (see Haberle et al., 2019; Kahre et al., 2020). Level-2 Mellor and Yamada is inherently diffusive and can cause thermal instabilities to develop in areas where radiative forcing is strong enough. A convective adjustment is performed every timestep to prevent such instabilities from growing (Haberle et al., 2019).

4.1.3 Radiative transfer.

The radiatively active atmospheric constituents in the model can include gaseous water vapor, gaseous CO$_2$, airborne dust, and water ice cloud aerosols, but clouds are not included in our simulations (Haberle et al., 2019). The radiative transfer (RT) code in the MGCM handles the radiative effects of these constituents using a 2-stream RT code (Kahre et al., 2020; Toon, Mckay, Ackerman, & Santhanam, 1989). Gaseous opacities for CO$_2$ and H$_2$O are calculated from correlated-K tables and then passed to the 2-stream code (Haberle et al., 2019). The optical properties of both dust and water ice cloud aerosols depend largely on their effective particle size distributions. These can be calculated from the aerosol mass and number mixing ratios that are predicted in the model and carried as observed quantities. The extinction efficiencies and scattering properties of dust, water ice, and other aerosols are calculated from Mie theory, and Rayleigh scattering from CO$_2$ is calculated directly (Bertrand et al., 2020). From these optical properties, the RT code generates fluxes and flux divergences that are used to calculate radiative heating and cooling rates throughout the atmosphere (Bertrand et al., 2020).

Dust radiative heating is the most influential diabatic forcing process in MGCMs, and the radiative effects of dust are particularly sensitive to the size distribution of airborne dust particles in the model (Haberle et al., 2019). Specifically, decreasing the dust effective radius ($r_{eff}$), which is the area weighted mean radius of the particle size
distribution, has the effect of increasing the visible-to-IR dust opacity ratio, which weakens the radiative-dynamic feedback of airborne dust (Murphy, Haberle, Toon, & Pollack, 1993). For example, in Bertrand et al. (2020), decreasing $r_{\text{eff}}$ from 3.0 µm to 1.5 µm halved the strength of the maximum mass flux in the simulated Hadley circulation.

4.1.4 Dust.

There are several dust lifting schemes that can be used in the MGCM. The simplest option initializes the model with a vertically-fixed dust distribution defined by a user-prescribed Conrath parameter. On the other end of the spectrum, there is an option allowing fully-interactive dust lifting in which model predicted surface stresses exceeding a critical value cause dust to be injected into the atmosphere. There is also a dust devil lifting parameterization scheme that is adapted from Rennó, Burkett, and Largin (1998) and modified for the Martian atmosphere. Finally, there is an option for injecting dust into the atmosphere according to observed column dust opacities retrieved from orbit. Hereafter called the “map-tracking scheme,” this is the option employed by the MGCM in this work.

The map-tracking scheme injects dust into the lowest model layer according to the abundance and distribution of dust in a series of dust absorption maps that are selected at model initialization. The dust absorption maps were created by Montabone et al. (2015). There are 11 MYs of dust absorption maps available, one for each year between MY24–MY34, and each year is represented by a set of 60 maps spaced every 6° of $L_s$. The MGCM assumes the maps vary linearly with time. Each map informs the model when, where, and how much dust to inject into the lowest model layer (Haberle et al., 2019). This dust lifting scheme allows the radiation, transportation, and sedimentation processes to occur with limited external influence which, in turn, allows the global
distribution of dust to evolve more realistically (Haberle et al., 2019). This method is also useful for reproducing specific dust events, such as the B storm, in specific MYs.

Dust injected into the lowest model layer has a fixed, log-normal size distribution defined by a user-assigned effective radius, $r_{eff}$. The nominal value of $r_{eff}$ in the MGCM is 2 µm. Once airborne, atmospheric dust can be transported by model winds and can fall out of suspension due to gravitational sedimentation, allowing the size distribution of the airborne dust in the atmosphere to evolve throughout the simulation (Haberle et al., 2019). This is important because the optical effects of the airborne dust depend on the size distribution of the dust. The evolving size distribution is represented by a two-moment scheme. The mean particle size of the dust is represented by independently-varying mass and number mixing ratios, and the effective variance of the dust has a fixed value of 0.5 (Bertrand et al., 2020; Haberle et al., 2019). The actual particle size distribution can therefore be calculated from the dust mass and number mixing ratios that are output by the model.

The rate at which airborne dust falls out of suspension depends primarily on the size of the suspended particles. The gravitational sedimentation rate (or fall velocity) is larger for sand-size particles than for small particles because gravity pulls more forcefully on larger particles (Kahre et al., 2017; Murphy et al., 1993). The fall velocity, $V_g$, of a particle of radius $r$ can be calculated using (Rossow, 1978):

$$V_g = \frac{2 \rho_p r^2 g}{9 \eta} (1 + \alpha K_n)$$  \hspace{1cm} (5)

where the dust particle density, $\rho_p$, is 2.5 kg m$^{-3}$, the gravitational constant for Mars, $g$, is 3.72 m s$^{-2}$, the atmospheric viscosity, $\eta$, is calculated from Sutherland's law, the Knudsen number, $K_n$, is given by $K_n = \frac{\lambda_R}{R_p}$, and $\alpha$ is given by

$$\alpha = 1.246 + 0.42 \exp \left( -0.87 \frac{0.87}{K_n} \right)$$  \hspace{1cm} (6)
Recall in Section 2.3.3 that fine particles that enter directly into suspension are ≤ 10 µm in diameter, and that saltating sand-size particles are ~ 100 µm to several hundred microns in diameter (Greeley & Iversen, 1985). Using equation 5, we calculate that the fall velocities of a fine-grain particle (4 µm radius), a coarse particle (10 µm radius), and a sand-size particle (100 µm radius) are ~ 0.3 cm s⁻¹, ~ 3.6 cm s⁻¹, and ~ 286.7 cm s⁻¹, respectively. To illustrate the difference between these fall velocities, consider a scenario in which the 100 µm particle and the 2 µm particle fall to the surface from an altitude of 5 m. From this height, it would take about ~ 1.74 seconds for the 100 µm sand-size particle to fall to the surface and nearly 2.3 hours for the 2 µm fine-grain particle to fall to the surface. This illustrates how fine-grain particles can be lofted tens of kilometers and carried several hundred thousand kilometers before falling out of suspension while sand-size particles that are lofted travel no more than a few meters at a time (Greeley & Iversen, 1985; Haberle & Leovy, 1982; Kahre et al., 2017).

4.2 Simulation setup.

We conducted several simulations with the NASA Ames MGCM at low resolution (4°x4°, or ~ 240x240 km) to determine the optimal settings for reproducing the B storm. The low resolution simulation was initialized with 28 vertical layers that were arranged in an altitude-varying pattern resulting in a vertical resolution of ~ 30 m near the surface to ~ 10 km near the model top (~ 90 km). Cloud microphysics were excluded from the simulations, dust injection was not permitted in areas covered by CO₂ ice, and the map-tracking dust lifting scheme tracked the MY31 dust absorption maps from Montabone et al. (2015). After running ten simulations with varying $r_{eff}$ between 1 µm–3 µm, we found that an $r_{eff}$ of 2.4 µm and 3.0 µm best reproduced the observed distribution of dust during the MY31 B storm. We chose to continue at high resolution using an $r_{eff}$ of 3 µm because the simulated temperatures in that simulation were in better agreement
with MY31 observations, and Bertrand et al. (2020) used an \( r_{\text{eff}} \) of 3 µm to reproduce the MY34 GDS in the MGCM.

Our baseline simulation was run at high resolution (1° x 1°, or \( \sim 60 \times 60 \text{ km} \)) with two additional layers in the vertical grid near the surface. The 30-layer vertical grid has a near-surface vertical resolution of \( \sim 6 \text{ m} \) and maintains the \( \sim 10 \text{ km} \) vertical resolution near the model top (\( \sim 90 \text{ km} \)). Model initialization was otherwise unchanged from the low resolution setup. The high-resolution baseline simulation ran for two MYs, using the MY31 dust map in the map-tracking scheme for both years, and our analysis was performed on model data output between \( L_x = 180°–360° \) in the second year of the simulation.

The MGCM ran on the Broadwell nodes of the Pleiades supercomputing system at NASA Ames Research Center. Each Broadwell node has two 14-core E5-2680v4 Intel Xeon processors and can run 28 processes of up to 128 GB of active memory in each node. At low resolution, the model ran in parallel using 72 processors spread across 3 Broadwell nodes (24 processing cores per node), and, at high resolution, the model ran in parallel across 30 Broadwell nodes and required all 28 processing cores in each node (840 processors total). The high-resolution simulation required \( \sim 18 \text{ hours} \) of walltime to complete a two-year simulation from a cold start.

### 4.3 Results

We present the results of the modeling analysis in this section, and we begin by comparing the MGCM-simulated and MCS-observed zonal mean thermal structure of the atmosphere during the southern summer season. We then compare the MGCM-predicted B storm temperatures and dust mixing ratios to the TES- and MCS-observed fields that were presented in Section 3.2. In Sections 4.3.4–4.3.6, we describe the MGCM-simulated...
mean state of the atmosphere, the dust lofting mechanisms involved in the simulated B storm, and the dust lifting mechanisms involved in lifting dust in the simulated B storm.

4.3.1 The zonal mean thermal structure of the MGCM-simulated atmosphere.

The zonal mean temperatures shown in Figure 18 are simulated (top) and MCS-observed (bottom) seasonal averages calculated over 30° of \( L_s \) around \( L_s = 270° \) (southern summer solstice). Figure 18 shows that the MGCM reasonably reproduces the thermal structure of the atmosphere during southern summer. Near-surface temperatures are maximized around \( > 240 \) K between 30° S and 90° S, and minimized around \( \sim 145 \) K at the north pole. The thermal gradient is largest near the surface in the northern hemisphere between 30°–70° N, and the northward-tilting warm tongue develops above 10 Pa between 65° N and the north pole.
Figure 18. Seasonal average zonal mean temperatures during the southern summer season on Mars. Data are averaged from $L_s = 255°–285°$. Contours are labeled inline every 10 K.

The MGCM differs from the observations primarily in the coldest regions of the atmosphere where the model predicts temperatures are as much as $\sim 20$ K colder than MCS observes. For example, in Figure 18, temperatures between 5–1 Pa are $\sim 140–160$ K in the simulation as opposed to $\sim 160–170$ K in the observations and, at the south pole, simulated temperatures cool more quickly with height than observed. At the south pole above $\sim 300$ Pa, temperatures are cooler than observed, which may impact the strength of the simulated B storm. Nevertheless, the MGCM captures the warm temperatures below 100 Pa in the southern hemisphere. Both simulated and observed temperatures are $> 320$K between the south pole and $\sim 20°$ N latitude, and the upward-curving pattern of
the lower half of the warm tongue (below $\sim 10$ Pa) is captured in the model. The MGCM also reproduces the cold region over north pole where temperatures are $\sim 150$ K north of 65° N and as cold as $\sim 140$ K poleward of 75° N throughout the middle atmosphere. In summary, the simulated southern summer atmosphere is well-represented in the model, especially the warmer regions between the surface and 5 Pa.

4.3.2 The B storm at 50 Pa.

In this section, we describe the MGCM-simulated 50 Pa temperature and dust fields as they evolve during the simulated B storm. First, the 50 Pa zonal mean temperatures and dust mixing ratios are shown in Figures 19 and 20, respectively. Selected comparisons are made to the observations that were shown in Figures 9 and 10 in Section 3.2.1, and then the simulated fields are shown in 10° $L_s$ intervals over the period from $L_s = 247°–287°$ in Figures 21 and 22. As in Figures 12 through 15, the data in Figures 21 and 22 are binned in 10° $L_s$ increments and the observed CO$_2$ cap edge is contoured.

Figures 19 and 20 show the simulated B storm develops in the high southern latitudes at the correct time of year. Simulated 50 Pa temperatures and dust concentrations are commensurate with those observed in the real B storm. Specifically, Figures 19 and 20 indicate that simulated 50 Pa temperatures exceed 200 K and simulated dust mixing ratios exceed 4 ppm over the highest southern latitudes around southern summer solstice ($L_s = 270°$), but that simulated B storm temperatures maximize around $\sim 10$ K cooler than observed and simulated dust mixing ratios maximize around $\sim 3$ ppm lower than observed. Warm B storm temperatures ($> 200$ K) and high dust mixing ratios ($> 4$ ppm) are sustained for $\sim 35°$ of $L_s$ in the simulation, and the simulated B storm peaks at $L_s = 267°$ as observed.
Figure 19. Simulated zonal mean 50 Pa temperatures over the period from $L_s = 230^\circ$–$290^\circ$. The red contour represents 200 K. The dashed line indicates the seasonal CO$_2$ cap edge from Piqueux et al. (2006).

Figure 20. As in Figure 19 but for simulated dust mixing ratios.

Although the simulated B storm is less intense than the observed B storm, the model captures its timing, location, and behavior very well. One of the key characteristics that the model captures is that the B storm dissipates more slowly than it initiates. In Figure 19, the 200 K isotherm can be seen tapering off after $L_s = 270^\circ$ which indicates the B storm decays slowly. Other characteristics captured in the simulation include the location of the maximum 50 Pa temperatures and dust mixing ratios (poleward of the observed receding CO$_2$ cap edge) and the lack of a northern hemisphere response in the 50 Pa temperature fields. Overall, the model captures the basic characteristics of the observed B storm and we are therefore confident that the MGCM can provide insight into the mechanisms responsible for dust lifting and lofting involved in the B storm. In the
following analysis, we explore the qualitative characteristics of the simulated B storm as it evolves from $L_s = 247^\circ - 287^\circ$.

To first order, Figures 21 and 22 show that the evolution of the simulated B storm resembles the evolution of the observed B storms. As observed, simulated 50 Pa temperatures are higher on the dayside than the nightside even though the south polar region experiences direct solar radiation at all hours of the day at this time of the year (Figure 21). Also as observed, the simulated dust field is more widely dispersed on the nightside than on the dayside, and dust mixing ratios maximize on the nightside (Figure 22).

During the growth phase of the simulated B storm ($L_s = 247^\circ - 257^\circ$), simulated 50 Pa temperatures over the seasonal CO$_2$ cap systematically increase by $\sim 10$ K every 10° of $L_s$ until peak intensity, and simulated 50 Pa dust mixing ratios increase by $\sim 4$ ppm over this period. Poleward of 80° S, 50 Pa dust mixing ratios increase from $< 2$ ppm at $L_s = 247^\circ$ to $\sim 6$ ppm at $L_s = 267^\circ$, and 50 Pa temperatures increase from $\sim 180$ K at $L_s = 247^\circ$ to $\sim 204$ K at $L_s = 267^\circ$. The 50 Pa temperature and dust fields maximize poleward of 80° S at peak intensity ($L_s = 267^\circ$).
Figure 21. Simulated 50 Pa temperatures at (top row) 3 PM and (bottom row) 3 AM for (columns) $L_s = 247^\circ, 257^\circ, 267^\circ, 277^\circ,$ and $287^\circ$. The data are binned in $10^\circ$ of $L_s$ increments. The red contour indicates the seasonal CO$_2$ cap edge from Piqueux et al. (2006).

Figure 22. As in Figure 21 but for dust mixing ratios. The black contour indicates the seasonal CO$_2$ cap edge from Piqueux et al. (2006).

As observed, the decay period of the simulated B storm is slow. At $L_s = 277^\circ$, more than half the dust has fallen out of suspension over the pole and corresponding 50 Pa
temperatures fall below 200 K. At $L_s = 287^\circ$, there is little change in the 50 Pa temperatures, but the simulated dust field displays some unusual behavior that is not observed by either TES or MCS. Specifically, Figure 21 shows that simulated 50 Pa temperatures and dust concentrations throughout the column poleward of $70^\circ$ S are slightly higher at $L_s = 287^\circ$ than at $L_s = 277^\circ$. This phenomena appears to be unique to the model since there is no evidence in the observations that suggests the increasing dust content at $L_s = 287^\circ$ is representative of the real atmosphere. This unusual behavior could be an effect of the map-tracking dust lifting scheme, but determining the cause of this organized dust pattern in the simulation is outside of the scope of this research.

4.3.3 Vertical cross-sections of the simulated B storm.

In this section, we present cross-sections of the temperature and dust fields in the simulated B storm by recreating the zonal mean temperature and dust cross-sections from Section 3.2.3 using simulated data. The simulated zonal mean temperature and dust mixing ratio cross-sections are shown in Figures 23 and 24. As in Figures 16 and 17, Figures 23 and 24 show zonal mean nightside (3AM) values to the left of the pole, zonal mean dayside (3PM) values to the right of the pole, and the data are binned by 10° of $L_s$ and shown in 10° of $L_s$ increments between $L_s = 247^\circ–287^\circ$.

Figure 23 shows that the thermal structure of the simulated B storm largely resembles that of the observed B storms especially during the onset and cessation periods. At $L_s = 247^\circ$, simulated temperatures are warmest in the mid-latitudes and cool throughout the column over the pole. The B storm grows warmer and deeper throughout the growth phase ($L_s = 247^\circ–257^\circ$) as the warm column of midlatitude air moves toward the south pole and as dayside 50 Pa temperatures approach 200 K. At the height of the B storm, $L_s = 267^\circ$, the warm column of air resides over the south pole and 50 Pa temperatures
exceed 200 K. The B storm dissipates quickly thereafter, and the warm (> 200 K) column of air can be seen growing shallower between $L_s = 277°–287°$.

**Figure 23.** As in Figure 16 but for simulated temperatures.

The simulated zonal mean dust cross-sections are shown in Figure 24. The simulated zonal mean dust injection rate is contoured in blue over the dust field and it is included to illustrate the relative location and amount of dust being injected into the lowest model layer during the B storm. Figure 24 shows that the model captures the pluming behavior seen in the B storms observed by MCS and shown in Figure 17. At $L_s = 247°$, a shallow dust column develops over $70°$ S between the surface and $\sim 50$ Pa (Figure 24). The associated dust injection rate is maximized at $65°$ S, just north of the developing plume, and some dust injection is occurring over the pole.

**Figure 24.** As in Figure 23 but for dust mixing ratios.
Over the next 20° of $L_s$, the dust plume travels poleward and grows deeper and dustier until it reaches the south pole at $L_s = 267°$. The associated dust injection rate increases and shifts poleward as well. At peak intensity ($L_s = 267°$), the dust column and the dust injection rate are maximized over the pole, and most of the simulated dust is confined to the 70° S latitude circle as seen in the observations. Dust mixing ratios are within $\sim 2$ ppm of the observed values in MY31 and exceed $\sim 4$ ppm below $\sim 25$ Pa. The simulated dust column is shallower and less dusty over the pole by $L_s = 277°$ as the B storm dissipates and dust falls out of suspension. At $L_s = 277°$, dust mixing ratios at 50 Pa are $\sim 2–3$ ppm lower than they were at $L_s = 267°$.

4.3.4 Simulated mean state of the atmosphere.

In this section, we describe the simulated zonal mean circulation during the southern summer season, and discuss how it may or may not be conducive to lofting dust in the B storm. The simulated zonal mean zonal wind, mass streamfunction, and meridional wind are shown in Figures 25, 26, and 27, respectively. The data are averaged over 30° of $L_s$ around $L_s = 270°$ as were the zonal mean temperatures shown in Section 4.3.1. Zonal and meridional winds are predicted by the model directly, and the mass streamfunction is a derived quantity calculated from the meridional wind field.

The simulated zonal mean zonal winds during southern summer, shown in Figure 25, feature a strong westerly jet in the northern hemisphere with wind speeds exceeding 130 m s$^{-1}$, and a weak easterly jet in the southern hemisphere with wind speeds exceeding $\sim 72$ m s$^{-1}$. The westerly jet is centered at 60° N and tilts poleward with height. The bottom of the jet is at $\sim 100$ Pa and creates strong vertical wind shear between the surface and 100 Pa. The easterly jet is centered over 10° S and produces its strongest winds between $\sim 10–3$ Pa. Above 100 Pa, there are easterly winds with speeds around $\sim 20$ m s$^{-1}$ throughout the southern hemisphere. The easterly jet tilts toward the south pole and
dominates most of the southern hemisphere above 100 Pa. Near the surface, weak westerlies are located at 30° S and 70° S below ~200 Pa.

Figure 25. Seasonal average ($L_s = 255°-285°$) zonal mean zonal winds during the southern summer season. Contours are labeled inline every 12 m s$^{-1}$.

The simulated zonal mean mass streamfunction, shown in Figure 26, captures the strong equator-crossing Hadley circulation characteristic of the Martian atmosphere during southern summer. The Hadley circulation has a strong rising branch in the southern hemisphere at ~40° S and a strong sinking branch in the northern hemisphere at ~60° N. The simulated Hadley cell produces a maximum mass flux greater than 100x10$^8$ kg s$^{-1}$ near the surface and the mass flux is strong throughout most of the column. Mass flux values of ~5x10$^8$ kg s$^{-1}$ are located as high as 10 Pa over the equator.

The strength of the Hadley circulation is reflected in the zonal mean meridional wind shown in Figure 27. The northerly winds maximize around ~30 m s$^{-1}$ at 10 Pa just north of the equator where the Hadley circulation accelerates the northward flow downwards, and the near-surface return flow can be seen over 20° S maximizing at > 10 m s$^{-1}$. 

80
Figure 26. Seasonal average ($L_s = 255^\circ$–$285^\circ$) zonal mean mass streamfunction during the southern summer season. Contours are labeled inline at the increments labeled in the colorbar.

The simulated zonal mean mass streamfunction shows a few smaller circulations poleward of the equator-crossing Hadley cell. South of the dominant Hadley cell is another thermally-direct Hadley cell (blue) between $40^\circ$ S–$60^\circ$ S, which is the remnant of the equinoctial southern Hadley cell shown in Figure 6. Between equinox and solstice ($L_s = 180^\circ$–$270^\circ$), the two-Hadley cell setup evolves into the structure seen in Figure 26 because of the strong pole-to-pole temperature gradient that develops around solstice and strengthens the northern Hadley cell. The small Hadley cell has a maximum mass flux $< 5 \times 10^8$ kg s$^{-1}$, which is less than one-twentieth the strength of the dominant Hadley cell. The other three cells in Figure 26 are, from left to right, a high-latitude thermally-direct cell (in blue) over the south pole, a weak thermally-indirect cell between $60^\circ$ S–$80^\circ$ S (in orange), and a thermally-indirect Ferrell cell over the high northern latitudes (in blue).
Figure 27. Seasonal average zonal mean meridional winds during the southern summer season. Data are averaged from $L_s = 255^\circ - 285^\circ$. Contours are labeled inline every 5 m s$^{-1}$.

The location of the rising branch of the Hadley cell is important to reproduce accurately because the vertical winds in that region are strong enough to loft dust tens of kilometers and transport it globally (Haberle, 1986; Haberle & Leovy, 1982). The rising branch of the Hadley cell in our simulation is at $\sim 40^\circ$ S, which is in agreement with the literature (see Barnes et al., 2017; Lewis, 2003; Mitchell et al., 2015), but this is too far north to interact with dust lofted in the B storm. Thus, the Hadley circulation is not involved in lofting dust in the B storm.

Since the rising branch of the Hadley cell is tens of degrees in latitude north of the south pole, the south polar environment is likely dominated by other, smaller circulations. Figures 25, 27, and 29 show that the high southern latitudes experience very weak synoptic-scale winds at this time of year. The magnitude of the zonal wind is less than $\sim 12$ m s$^{-1}$ poleward of $70^\circ$ S throughout most of the lower and middle atmosphere. The meridional wind speed is $< 1$ m s$^{-1}$ near the surface at the south pole, and the vertical wind speed is $\sim 1 - 3$ cm s$^{-1}$ over the south polar cap. Warm temperatures and low prevailing winds provide ample opportunity for local circulations to drive dust lifting and lofting during the B storm.
There is little baroclinic wave activity at the south pole at this time of year because baroclinic instabilities require large horizontal temperature gradients, strong vertical wind shear, and relatively flat topography (Barnes et al., 1993; Lewis, 2003; Pollack et al., 1990). However, there is a sea breeze-like circulation over the south pole called the cap-edge breeze that is likely a major player in lifting dust but is unlikely lofting dust in the B storm because it causes air to sink over the pole.

The simulated zonal mean temperatures, vertical winds, and meridional winds averaged over $L_s = 252°–262°$ (during the B storm growth phase) are shown in Figures 28 and 29 to illustrate the strength of the simulated cap-edge breeze. The mean and maximum latitudinal extent of the observed CO$_2$ cap edge are labeled $\phi_{\text{mean}}$ and $\phi_{\text{max}}$, respectively. Figure 28 shows that the temperature gradient increases from $\sim 237$ K over the cold CO$_2$ ice at 90° S to $\sim 249$ K over the warm dry surface at 71° S ($\phi_{\text{max}}$). The thermal gradient over the cap produces a temperature change of $\sim 14$ K across the southernmost $< 30°$ of latitude. This produces a circulation in which air rises at $\sim 1$ cm s$^{-1}$ over the cap edge ($70°–80°$ S), sinks at $\sim 3$ cm s$^{-1}$ over the south pole, and produces low-level southerly winds at the surface that may be strong enough to lift dust along the cap edge (Figure 29).
**Figure 28.** Zonal mean temperatures averaged over $L_s = 252^\circ - 262^\circ$ during the growth phase of the B storm. Contours are labeled inline every 2 K. The mean and maximum latitudinal extent of the observed seasonal CO$_2$ ice cap are labeled $\phi_{\text{mean}}$ and $\phi_{\text{max}}$.

**Figure 29.** As in Figure 28 but for simulated vertical winds.
4.3.5 Dust lofting.

In Section 2.3.5, rocket dust storms and the solar escalator effect were two local-scale processes introduced as potential dust lofting mechanisms. In this section, we use the MGCM to look at the hour-by-hour development of the simulated B storm to determine whether these processes might be lofting dust in the B storm and, if they are, where they develop and how they operate. We examine the instantaneous dust field at hourly increments over a two-day period during the peak of the B storm, $L_s = 269.54^\circ - 270.64^\circ$. We highlight seven hours within the 48-hour period that illustrate the evolution of the dust lofted during the simulated B storm. The data presented in Figures 30–38 are instantaneous cross-sections of dust concentrations, shortwave heating rates, dust injection rates, and vertical wind speeds at 70° S and 75° S latitude. Shortwave heating rates are contoured over the dust field, arrows indicate the magnitude and direction of the vertical velocity field, and the vertical dashed line indicates the longitude of local noon. We have subtracted the dust particle fall velocities from the vertical wind field so that the vector field represents the net vertical velocity of the local environment. The rolling average dust injection rate, calculated over the preceding 20° of $L_s$, is shown below the cross-section to illustrate where dust injection is occurring.

Figure 30 shows dust mixing ratios, vertical winds (arrows), and local heating rates (contours) at $L_s = 269.54^\circ$ across 70° S latitude. Local noon is at 180° E and a westward tilting dust plume is developing over 130° E. Throughout the 48-hour period, the plume develops in the eastern hemisphere and the dust injection rate maximizes in the eastern hemisphere. The most optically thick part of the plume extends from the surface to $\sim$ 100 Pa, almost two scale heights, and mixing ratios are $> 17$ ppm where heating rates are $> 70$ K sol$^{-1}$. Weak positive vertical velocities are associated with the dusty western edge of the plume at $\sim$ 100° E, and some dust is present between 100–50 Pa in smaller
concentrations (∼ 10 ppm) and associated with local heating rates of 30–45 K sol⁻¹. Smaller dust mixing ratios (∼ 4–5 ppm) are present at 50 Pa, the level where the B storm is defined, over 180° W and 60° E.

Figure 30. Simulated dust mixing ratio cross-section along 70° S at Lₘ = 269.54, the first hour in the series. Heating rates (K sol⁻¹) are contoured in black and white and labeled inline. Vertical velocities are represented by arrows that indicate the magnitude and direction of flow. Blue arrows represent rising motion and indigo arrows represent sinking motion. The vertical dotted line indicates the local noon and may not be visible when local noon is over 180°.

Figure 31 shows the dust field nine hours later at Lₘ = 269.78°. The thickest part of the plume is located at the latitude of local noon, ∼ 50° E, and has mixing ratios of up to ∼ 17 ppm throughout the lowest 2.5 scale heights. Dust mixing ratios at 50 Pa are ∼ 6 ppm over most longitudes and heating rates are > 30 K sol⁻¹ throughout most of the
eastern hemisphere. The upper part of the plume is beginning to detach from its lower half at \( \sim 150 \) Pa. Local heating rates maximize at \( > 60 \) K sol\(^{-1} \) and produce strong vertical velocities in the dustiest areas over \( 40^\circ \) E and \( 100^\circ \) E. Positive vertical velocities are present throughout the lowest two scale heights of the eastern hemisphere and associated heating rates are at least \( 30 \) K sol\(^{-1} \) throughout the region as well.

**Figure 31.** As in Figure 30 but for hour 10.

By the fourteenth hour (\( L_s = 269.89^\circ \)), shown in Figure 32, the plume extends well into the western hemisphere and dust is being lofted as high as \( 15 \) Pa between \( \sim 30^\circ - 50^\circ \) W. Dust mixing ratios at \( 50 \) Pa are \( > 4 \) ppm over most longitudes and \( > 6 - 8 \) ppm within the plume over \( 30^\circ \) W. Associated heating rates are \( > 45 \) K sol\(^{-1} \) and vertical velocities are positive where mixing ratios are greatest. The upper part of the plume extends from
\( \sim 100-15 \text{ Pa}, \sim 2-3 \text{ scale heights, centered over } \sim 20^\circ \text{ W} \) which is also the latitude of local noon. Positive vertical velocities throughout the upper half of the plume indicate air is rising faster in the upper part of the plume than in the lower part of the plume. In the lowest 1.5 scale heights, mixing ratios between 25°–100° E are > 17 ppm and associated heating rates are > 60 K sol\(^{-1}\).

Figure 32. As in Figure 30 but for hour 14.

Five hours later at \( L_s = 270.02^\circ \) (Figure 33), the plume has split. A detached dust layer resides between 25–7 Pa at 140° W (just a few hours before local noon) and has mixing ratios of \( \sim 7 \text{ ppm} \) and heating rates greater than 30 K sol\(^{-1}\). There is less dust below the detached layer at 50 Pa where mixing ratios are \( \sim 2-4 \text{ ppm} \) everywhere except over 25° W. At 25° W, a second mid-level detached dust layer with heating rates > 30 K
sol$^{-1}$ produces strong rising motion. In the lowest two scale heights, the eastern hemisphere is very dusty with mixing ratios $> 17$ ppm and heating rates $> 60$ K sol$^{-1}$.

**Figure 33.** As in Figure 30 but for hour 19.

At the 25th hour, $L_s = 270.19^\circ$, Figure 34 shows the detached layer re-enters the eastern hemisphere over 150° E between 25–5 Pa. Dust mixing ratios are $\sim 7$ ppm in the center of the detached layer and heating rates are as high as 45 K sol$^{-1}$. Strong vertical velocities are associated with the detached dust layer. Meanwhile, the surface dust plume that produced the detached layer is redeveloping over 75° E. Dust mixing ratios are $\sim 17$ ppm throughout the lowest two scale heights and mixing ratios are $\sim 10$ ppm up to and above 50 Pa. Heating rates are $> 45$ K sol$^{-1}$ throughout most of the plume and are $> 30$ K sol$^{-1}$ at and above 50 Pa. Positive vertical velocities are present throughout the plume.
The redevelopment of the plume shown at 70° S in Figure 34 is more clearly defined 5° further south at 75° S, which is shown in Figure 35. In Figure 35, the dust plume extends from the surface to 5 Pa, nearly five scale heights, and has mixing ratios > 7 ppm and associated heating rates > 30 K sol\(^{-1}\). The uppermost part of the plume at 10°–20° W is about 12 hours removed from local noon which means the plume is developing around local midnight. At 40° E between 75–15 Pa dust concentrations are > 10 ppm, heating rates are > 45 K sol\(^{-1}\), and positive vertical velocities are strong. Between 15–5 Pa over 10° W in the center of the plume, mixing ratios are > 14 ppm and heating rates are as high as > 60 K sol\(^{-1}\). The detached dust layer can also be seen between 25–5 Pa at
Figure 35. As in Figure 30 but for hour 25 at 75° S.

The well-defined structure of the plume in Figure 35 is a good example of the strength and depth of the plumes that form in the simulated B storm. The dust particle size distribution, sedimentation rates, vertical winds, and dust mixing ratios associated with the plumes shown in Figure 35 are displayed in Figure 36. Figure 36 shows that the core of the dust plume is made up of dust particles with radii around 2-2.5 µm whose fall velocities, calculated using Equation 5, are on the order of 0.3-0.35 cm s$^{-1}$. The vertical velocity field in Figure 36 indicates that the upward winds associated with the uppermost part of the plume are on the order of $\sim 10$ cm s$^{-1}$, which is more than 30 times greater.
than the fall velocities of the 2.5 µm dust particles. As a result, the airborne dust in the plume remains in suspension because the local environment is experiencing net upward velocities of \( \sim 9.7 \text{ cm s}^{-1} \).

**Figure 36.** From left to right: dust mixing ratios and associated heating rates (as in Figure 35), dust particle radii, sedimentation rates, and vertical velocities for hour 25 at 75° S.

At hour 36 \((L_s = 270.48°)\) in Figure 37, the detached dust layer has dissipated entirely and a new dust layer has detached from the plume that was redeveloping in Figures 34 and 35. The detached layer is located over 140° W between 25–3 Pa and has almost twice as much dust as the first layer had when it was at the same longitude in Figure 33. The detached layer has mixing ratios \( > 10 \text{ ppm} \) and associated heating rates of \( > 45 \text{ K sol}^{-1} \). The surface plume is tall and optically thick with mixing ratios \( > 17 \text{ ppm} \) stretching over 2.5 scale heights. The plume originates near the surface at 100° E and stretches to 100 Pa over 20° W. The highest part of the plume reaches 50 Pa over 20° E, which is also the longitude of local noon, and dust mixing ratios are \( \sim 6 \text{ ppm} \) throughout most of the atmosphere at 50 Pa. Between 75° W and 75° E, heating rates associated with the 50 Pa dust are as high as \( \sim 30 \text{ K sol}^{-1} \).
Figure 37. As in Figure 30 but for hour 36 at 75° S.

The surface plume continues to loft dust to $\sim 50$ Pa through the 48th hour. Detached dust layers form periodically but become less dusty and produce lower heating rates after solstice, $L_s = 270^\circ$. The 48th hour of the period ($L_s = 270.81^\circ$), shown in Figure 38, shows the surface plume is located west of its previous location, around 50° E, and it no longer tilts westward with height. The plume is once again developing around local midnight. Dust is lofted to 25 Pa and heating rates are $> 30$ K sol$^{-1}$ throughout the plume, but vertical velocities are no longer positive in the dustiest areas. There is, however, rising air associated with a detached dust layer between 50–15 Pa over 170° E and some weak positive vertical velocities are located within a detached layer between
25–5 Pa over 100° W. Between the two detached dust layers and the surface plume, there is still a significant amount of dust at 50 Pa at this time (L_s = 270.81°).

Figure 38. As in Figure 30 but for hour 48 at 75° S.

Several patterns emerge in the series of cross-sections presented here. First, shortwave heating rates are highly correlated with airborne dust. Second, the areas in which heating rates are highest are typically associated with positive vertical velocities. Third, dust is lofted above the boundary layer by localized pluming events that can develop independent of the longitude of local noon. Together, these suggest that dust lofting occurs in an environment made positively buoyant by the localized heating of suspended dust particles in the atmosphere. To test this theory, we perform a simulation designed to test the
sensitivity of the pluming mechanism to the radiative heating of the airborne dust in the model. The results of this sensitivity case are presented in the following section.

### 4.3.5.1 Sensitivity case: Lofting radiatively inert dust.

To test the theory that the radiative heating of airborne dust is the mechanism lofting dust in the B storm, we ran the baseline simulation with the radiation code decoupled from the transported dust. This was done by prescribing a dust scenario that has a fixed vertical distribution of dust defined by a Conrath parameter of 0.003 to the radiation code (although the vertical distribution of dust is fixed, the horizontal distribution of dust is still dictated by the MY31 dust map). The transported dust is forced to be radiatively inert, but it is injected into the atmosphere according to the map tracking scheme and it can be transported by model winds or fall out of suspension. Decoupling the radiation code from the transported dust allows us to determine what role dust radiative effects have on lofting in the B storm and whether, or to what extent, other dynamical and physical processes might be involved.

We present the results of the sensitivity case in Figure 39 which shows the 25th hour in the sequence ($L_s = 270.19$) at 75° S as in Figure 35. Only one hour from the sensitivity case is shown for the sake of brevity because there is little change in the dust activity, heating rates, or vertical winds throughout the period. Figure 39 shows most of the airborne dust is confined to the lowest scale height in the atmosphere ($< 250$ Pa) and dust concentrations in this region are $< 1$ ppm. There is no pluming and very little dust lofting in the sensitivity case and the atmosphere is nearly dust-free above one scale height. Heating rates are maximized just after local noon (180° E) at $\sim 60$ K sol$^{-1}$ near the surface and $> 30$ K sol$^{-1}$ throughout the first three scale heights, and positive vertical velocities are located throughout the lowest four scale heights in the eastern hemisphere.
where heating rates are maximized. Heating rates and vertical velocities are minimized
around local midnight (∼ 0° E).

Figure 39. As in Figure 35 but for the sensitivity case.

Atmospheric heating drives vertical motion in both the baseline simulation and the
sensitivity case. Strong, positive vertical velocities are co-located with strong heating
(> 30 K sol⁻¹) in both the baseline and sensitivity simulations (Figures 35 and 39), but
the radiative forcing mechanism is different. In the baseline case, the magnitude and
distribution of radiative heating depends primarily on the distribution of airborne dust
(Figure 35). With the radiative effects of dust removed, heating rates in the sensitivity
case are diurnally forced, maximizing at the longitude of local noon (Figure 39).
The largest positive vertical velocities in Figure 39 are located above $\sim 100$ Pa (about one scale height) which is too high to loft dust from the PBL to the free atmosphere. Some small positive vertical velocities exist near the surface at $100^\circ$ E and $150^\circ$ W, but little dust is present ($< 1$ ppm) above $\sim 250$ Pa at these longitudes. Since little airborne dust is maintained in suspension, the model injects dust at higher rates in an attempt to match the MY31 dust map. For example, the dust injection rate is maximized at $\sim 20^\circ$ W in Figure 39 where the uppermost part of the dust plume should be according to Figure 35.

The sensitivity test confirms that the localized heating of suspended dust particles is crucial for lofting dust in the B storm. In both the baseline and sensitivity cases, heating rates $> 30$ K sol$^{-1}$ cause air to rise but whether dust is lofted within the buoyant air depends on the radiative properties of the dust itself and is independent of the subsolar point. In the baseline case, dust radiative heating has greater influence on local heating rates than direct radiative heating from the sun. For example, in Figure 35, heating rates as high as $\sim 60$ K sol$^{-1}$ are located four scale heights above the surface at local midnight in a region of dusty air. In the sensitivity case, radiative heating from the sun drives local heating rates but dust is not lofted above $\sim 150$ Pa and heating rates rarely exceed 45 K sol$^{-1}$ above two scale heights ($\sim 100$ Pa; Figure 39).

4.3.6 Dust lifting.

The map tracking method of dust injection used in these simulations was described in detail in Section 4.2. This method is not an interactive dust lifting scheme, that is, the model does not lift dust where near-surface winds and surface stresses exceed a threshold value. Instead, dust is injected into the atmosphere only if the opacity in the MY31 dust map exceeds the model-transported dust opacity at the corresponding gridpoint. If these conditions are met, dust is injected into the lowest model layer at the relevant time and location regardless of the aeolian processes predicted by the model. In this sense, model
predicted surface stresses and dust injection rates are decoupled. As a result, we can diagnose how well the model is capturing dust lifting in the B storm by comparing the regions of high surface stress with the regions of high dust injection rates. In this section, we explore the physical processes that could be responsible for lifting dust in the B storm by comparing the MGCM-predicted surface stresses and dust injection rates.

Model predicted surface stresses and dust injection rates at the south pole are shown in Figure 40 for the period between $L_s = 237°–277°$. The data are binned by 10° of $L_s$, the observed seasonal CO$_2$ cap is outlined in white, and the fluid threshold (22.5 mN m$^{-2}$) as defined by Kahre et al. (2006) is contoured in black. Comparing the two fields reveals that the regions in which dust is injected into the atmosphere are not regions where high surface stresses are predicted. For example, dust injection increases between 65–70° S latitude from $L_s = 237–247°$, but surface stresses are maximized further south and weaken during this period. Surface stresses at $L_s = 237°–247°$ are maximized along the seasonal CO$_2$ cap edge but dust injection rates are not.

Figure 40. Simulated dust injection rates (top row) and surface stresses (bottom row) at (columns) $L_s = 237°, 247°, 257°, 267°, \text{ and } 277°$. The white contour indicates the seasonal CO$_2$ cap edge as observed by Piqueux et al. (2006) and the black contour indicates regions exceeding the fluid threshold for lifting sand-sized dust particles (22.5 mN m$^{-2}$).
Surface stresses at $L_s = 237^\circ$ are likely high enough to lift dust at several locations, which are outlined in black, and may be high enough to lift dust at $L_s = 247^\circ$ if the fluid threshold is just under 22.5 mN m$^{-2}$. However, dust lifting is unlikely to occur at $L_s = 257^\circ$. This is problematic because the B storm is observed developing (and presumably lifting dust) at $L_s = 257^\circ$ and the dust injection rate indicates dust should be lifted there. Dust injection rates are maximized at $\sim 65$ mg m$^{-2}$ per hour along 70° S in the western hemisphere at $L_s = 257^\circ$, yet the corresponding surface stresses are negligible. At the height of the B storm, $L_s = 267^\circ$, dust injection rates are lower, maximizing around $< 40$ mg m$^{-2}$ per hour, but surface stresses are still not high enough to lift dust.

The co-location of the maximum surface stresses and the observed CO$_2$ cap edges at $L_s = 237^\circ$ and $L_s = 247^\circ$ suggest cap-edge processes play a significant role in dust lifting during the B storm. However, the disappearance of significant surface stresses at $L_s = 257^\circ$ indicates the seasonal CO$_2$ ice cap sublimates more quickly in the simulation than observed. This is illustrated in Figure 41 which shows the model predicted, 5-sol average surface CO$_2$ ice mass every 10° of $L_s$ between $L_s = 237^\circ$–277°. Overlaid in white is the observed seasonal CO$_2$ cap edge.

Figure 41. Simulated CO$_2$ ice mass at (column) $L_s = 237^\circ$, 247°, 257°, 267°, and 277°. The white contour indicates the seasonal CO$_2$ cap edge as observed by Piqueux et al. (2006). The nominal cap albedo is used in the baseline simulation (0.51).

Figure 41 confirms that the surface CO$_2$ ice in the model sublimates almost entirely by $L_s = 257^\circ$, well before the observed ice disappears, and that the location of the
simulated ice is more symmetric around the south pole than observed. The simulated cap edge is roughly aligned with the observed cap edge at \( L_s = 237^\circ \), but the mass of surface CO\(_2\) ice is maximized in the eastern hemisphere opposite the observed CO\(_2\) ice cap. This offset is more prominent at \( L_s = 247^\circ \) when the observed cap is offset toward the western hemisphere and the simulated cap is offset toward the eastern hemisphere. The local minimum surface CO\(_2\) at 90° S is caused by the high thermal inertia region in the sourced map which is intended to represent the perennial CO\(_2\) ice cap. In the simulation, a small amount of surface CO\(_2\) ice remains at \( L_s = 257^\circ \) and all surface CO\(_2\) ice is gone by \( L_s = 267^\circ \), whereas the observed seasonal CO\(_2\) cap is present throughout most of the 80° S latitude circle at \( L_s = 257^\circ \) and sublimates slowly through \( L_s = 277^\circ \).

The early disappearance of the simulated seasonal CO\(_2\) cap precludes surface stresses from maximizing near the pole during the simulated B storm. Since the observed CO\(_2\) cap is still present from \( L_s = 257^\circ – 277^\circ \) and since the simulated cap edge and the simulated surface stresses appear to be correlated, we hypothesis that if the simulated cap persists as long as the observed cap, then the surface stresses will likely maximize along the cap edge throughout the B storm. In the next section, we present the results of another sensitivity test that lends merit to this theory.

4.3.6.1 Sensitivity case: Lifting dust along the southern seasonal CO\(_2\) cap.

To determine the sensitivity of the simulated surface stress to the location of the cap edge, we ran a simulation in which the albedo of CO\(_2\) ice at the south pole was raised from the nominal value of 0.51 to 0.8 to force the CO\(_2\) ice to sublimate more slowly. The simulated surface CO\(_2\) ice mass from the sensitivity case and the observed CO\(_2\) cap edge are shown in Figure 42. As in Figure 41, Figure 42 shows the 5-sol average surface CO\(_2\) ice mass at 10° of \( L_s \) intervals from \( L_s = 237^\circ – 277^\circ \).
Figure 42 shows that raising the south polar CO$_2$ ice albedo causes the surface CO$_2$ ice to linger $\sim 20^\circ$ of L$_s$ longer than in the baseline simulation. Surface CO$_2$ ice persists through L$_s = 277^\circ$ in the sensitivity case and there is better agreement between the model predicted cap edge and the observed cap edge from L$_s = 237^\circ$–257$^\circ$. As in the baseline case, the CO$_2$ ice cap in the sensitivity simulation is centered over the south pole and the surface CO$_2$ ice maximum is in the hemisphere opposite the observed CO$_2$ ice cap. The simulated cap edge is within a few degrees of latitude of the observed cap edge from L$_s = 237^\circ$–257$^\circ$ in Figure 42.

Figure 42. As in Figure 41 but for a simulation with a southern seasonal CO$_2$ ice albedo of 0.8.

At L$_s = 257^\circ$, the surface area covered by CO$_2$ ice is larger in the simulation than the area of the observed CO$_2$ ice cap, but this is an improvement over the baseline case in which only a sliver of ice remains in the eastern hemisphere at L$_s = 257^\circ$. At L$_s = 267^\circ$, some surface CO$_2$ remains in the eastern hemisphere at 80° S latitude, and the simulation captures the mean latitudinal extent of the cap edge at L$_s = 267^\circ$. Finally, only a small patch of surface CO$_2$ ice remains in the eastern hemisphere at L$_s = 277^\circ$ and very little persists at L$_s = 287^\circ$.

Simulated surface stresses and dust injection rates from the sensitivity case are shown in Figure 43. As in Figure 40, the data in Figure 43 are 5-sol averages shown every 10° of L$_s$ from L$_s = 237^\circ$–277$^\circ$. Figure 43 shows surface stresses in the sensitivity case are
maximized along the cap edge through $L_s = 267^\circ$, which is $20^\circ$ of $L_s$ longer than in the baseline case. The rate and location of dust injection are similar in both cases, but dust injection rates are higher and shifted slightly northward in the sensitivity case at $L_s = 257^\circ$ and $L_s = 267^\circ$. Injection rates are likely higher because the model cannot lift dust over areas covered in ice and therefore must lift the same amount of dust over a smaller area. The northward shift better reflects the location where the MGCM predicts high dust injection rates and surface stresses are occurring from $L_s = 257^\circ$–$267^\circ$. The surface stress maxima at $L_s = 257^\circ$ exceed the fluid threshold for lifting sand-sized particles from Kahre et al. (2006) at three locations around the cap edge: $\sim 0^\circ$ E, $\sim 80^\circ$ W, and $\sim 180^\circ$ E.

Figure 43. As in Figure 40 but for a simulation with a southern seasonal CO$_2$ ice albedo of 0.8.

A CO$_2$ ice albedo of 0.8 may not be unusually high for the seasonal CO$_2$ ice cap at this time of year. There is evidence that the southern seasonal CO$_2$ ice cap albedo varies throughout the lifetime of the seasonal cap, and that such variations affect the rate and location of CO$_2$ deposition and sublimation at the poles (Kahre et al., 2006). James, Bonev, and Wolff (2005) measured the brightness of the seasonal cap during its retreat
phase in MY26 and found the albedo of the CO$_2$ ice was 0.74 at $L_s = 235^\circ$, 0.87 at $L_s = 251^\circ$, and 0.67 at $L_s = 265^\circ$. MCS observations show the seasonal cap is brightest during the retreat phase and darkest during the growth phase, producing an average seasonal CO$_2$ ice albedo of 0.51 (Gary-Bicas et al., 2020). Prescribing this value to the southern seasonal CO$_2$ cap in the MGCM reproduces the seasonal pressure cycle observed by the Viking landers reasonably well and it is therefore the nominal CO$_2$ ice albedo for the south seasonal cap in the MGCM.

The sensitivity case illustrates that raising the seasonal CO$_2$ cap albedo from 0.51 to 0.8 slows the rate of sublimation of the CO$_2$ ice and produces a surface ice distribution that is in better agreement with the observations than the baseline case. Extending the lifespan of the simulated cap also produces surface stresses that are co-located with high dust injection rates during the B storm. Combined with the magnitude of surface stresses at $L_s = 257^\circ$, which exceed some of the higher fluid thresholds for lifting sand-sized particles, the results of sensitivity case suggest that cap-edge processes around the sublimating CO$_2$ cap could produce surface stresses high enough to lift dust at the south pole during the B storm.

In this section, we provided a description of the NASA Ames MGCM and the relevant physics packages that were employed in our simulations. We described the various dust lifting schemes available in the MGCM, placing emphasis on the dust map-tracking scheme used in our baseline simulation, and included a summary of how the simulated dust field is allowed to evolve after it is lifted. We then compared the MGCM-simulated zonal mean thermal structure of the atmosphere to MCS observations to show that our baseline simulation reproduced the observed atmosphere, and we highlighted where the MGCM reproduced the key characteristics of the B storm outlined in Section 3.2. The zonal mean state of the simulated atmosphere was presented to show that large-scale
circulations could not be responsible for lofting dust in the B storm and that local-scale processes are likely driving this process. With the high-resolution simulation, we were able to look at these local processes more closely to show that the radiative heating of suspended particles in the atmosphere is important for lofting dust in the B storm, and that lifting along the seasonal CO$_2$ cap edge is essential for reproducing the B storm in the MGCM.

In the following section, we discuss how the relationship between dust particle size, heating rate, and sedimentation rate might influence lofting in the simulated B storm. The interaction of these processes is significant in the solar escalator effect, and we discuss whether and to what extent our simulations indicate that the solar escalator effect is occurring in the B storm. Finally, we explore the mechanisms that could be involved in cap-edge lifting and how we intend to improve the simulated seasonal CO$_2$ cap retreat in order to test the sensitivity of these mechanisms to the location of the cap edge.
5 Discussion

The MGCM captures key aspects of the B storm reasonably well. As in the observations, the simulated B storm occurs over the highest southern latitudes around the southern summer solstice season ($\sim L_s = 245^\circ$–$290^\circ$). It peaks in intensity just before southern summer solstice at $L_s = 267^\circ$. B storm temperatures exceed 200 K and dust mixing ratios exceed 4 ppm at 50 Pa throughout the storm. These results are promising and the logical next step would be to simulate the B storm interactively. That is, to allow the lifting, transport, and sedimentation of radiatively active dust to be self-consistently determined. However, the MGCM-simulated B storm is less dusty than observed, and this causes the simulation to under-predict the temperature, intensity, and duration of the storm. Further investigation is required to address various questions about dust and dust storms that these discrepancies raise before we can expect to simulate the B storm using a fully interactive dust lifting scheme.

The MGCM predicts that near-surface particles with $\sim 2.5 \mu m$ radii are lofted tens of kilometers during the B storm, which raises questions regarding the nature of the balance between the radiative and physical properties of airborne dust on Mars. As discussed in Section 2.3.5, the radiative-dynamic feedbacks of airborne dust are sensitive to the size, local concentration, and the vertical distribution of particles in the atmosphere. Although larger particles (e.g. a few hundred microns in diameter) are more effective than smaller particles (e.g. $\leq 10 \mu m$) at heating the atmosphere, they also fall out of suspension faster (Haberle et al., 2019; Murphy et al., 1993). In our simulation, the MGCM lifts a size distribution of particle defined by an $R_{eff}$ of 3 $\mu m$ because this best reproduces the observed 50 Pa dust mixing ratios and temperatures during the B storm. However, the simulated B storm is cooler, weaker, and shorter in duration than observed and the size distribution of the lifted particles may be impacting the strength of the simulated B storm.
Additionally, the particles in the MGCM are spherical, which is likely an unrealistic assumption that may need to be re-examined.

The solar escalator effect, described in Section 4.3.5, occurs when localized heating of suspended dust in the atmosphere creates a buoyant environment in which airborne dust is continually lofted (Daerden et al., 2015). In our simulations, the co-location of dust, high heating rates, and positive vertical velocities within the eastern hemisphere dust plume resembles the solar escalator effect described in Daerden et al. (2015). We showed in Section 4.3.5 that the dust layers that detach from the plume typically travel one full rotation around the 70°–75° S parallel before dissipating, similar to the trajectory of the dust layer studied in Daerden et al. (2015). However, unlike in Daerden et al. (2015), the pluming behavior of the dust in our simulation does not consistently maximize at local noon. In fact, there are several plumes that form around local midnight, which calls into question whether or not the solar escalator effect is occurring in our simulations. Further analysis is needed to determine the trajectory of the lofted dust and to confirm that the solar escalator effect is indeed occurring in the B storm.

As the seasonal CO$_2$ ice cap sublimates during southern summer, the latitudinal surface temperature gradient over the highest southern latitudes increases and causes a sea breeze-like circulation in which sinking air over the cool polar ice and rising air over the warm dry surface causes near-surface northward winds. The cap-edge breeze, augmented by katabatic winds, the CO$_2$ sublimation flow, and the Coriolis force, is likely producing high surface stresses along the cap edge during this season. However, the simulated cap edge in our baseline simulation is not co-located with the observed cap edge nor does it sublimate at the same rate as the observed ice cap. In our sensitivity study (Section 4.3.6.1), we showed that the user-defined CO$_2$ ice albedo influences the retreat rate but does not change the retreat pattern. The retreat pattern of the observed seasonal CO$_2$ ice
cap is likely influenced by the residual CO$_2$ ice beneath it, but the MGCM does not reproduce the residual ice cap explicitly. Instead, the residual cap is represented by the regions of high thermal inertia and surface albedo, and the simulated surface CO$_2$ ice appears to be more sensitive to the thermal inertia of the surface than to its albedo. Further investigation is required to determine how to best represent the retreat of the seasonal CO$_2$ ice cap in the model and to confirm that dust is still lofted to 50 Pa during the B storm if the cap retreat rate and pattern are changed.

There is much to be explored in regards to dust and dust storms in the MGCM, and future investigations regarding the behavior of dust in the model are required before simulating the B storm interactively. The relationships between particle size, heating rate, and sedimentation rate are complex. Improving our understanding of how these processes operate could illuminate how the dust lofting mechanism operates, and such tests could confirm whether the solar escalator effect is occurring. The simulated seasonal CO$_2$ cap retreat rate and pattern need to be improved, and the way the model represents the residual CO$_2$ ice cap may need to be reevaluated. Addressing these issues is complicated but necessary in order to simulate the B storm interactively. We are optimistic that we have the tools required to meet these challenges in the near future.
6 Conclusion

The first goal of this research was to characterize the development and evolution of the annually-recurring regional B dust storm on Mars by performing an observational analysis of the temperature and dust fields retrieved from orbit. The second goal of this research was to identify and define the mechanisms lifting and lofting dust during the B storm. In this conclusion, we provide a summary of the results of the observational and modeling components of this work.

To date, the literature regarding the B storm is sparse. The B storm is defined by the 50 Pa temperature fields from TES and MCS in Kass et al. (2016) but is otherwise a small part of broad, descriptive analyses of dust activity on Mars. Part of the reason for this is that the B storm is unique compared to its A and C storm counterparts, but the B storm is also difficult to study because it is located in the highest southern latitudes where observations are limited and modeling is historically difficult. The development of the new NASA Ames MGCM provides an opportunity to simulate a polar phenomenon at high resolution and on a uniform polar grid, and the B storm is an ideal candidate for exploring the capabilities of this new model and for expanding our understanding of dust storm development on Mars.

The observational analysis of TES- and MCS-observed 50 Pa temperatures and dust mixing ratios show that in the years lacking a GDS between MY24 and MY33, an annually-recurring regional-scale B storm occurs at the south pole of Mars. It occurs over 30°–40° of L$_s$ and peaks just before southern summer solstice (L$_s$ = 270°). Temperatures at 50 Pa exceed 200 K, dust mixing ratios at 50 Pa exceed 4 ppm, and column dust opacities exceed 0.4 during this time. The majority of the mid-level dust is confined south of 60° S and temperatures > 200 K are confined south of 50° S. The mid-level dust tends to concentrate in the eastern hemisphere between 45°–180° E and temperatures maximize
at ~ 220 K in this region. Dayside temperatures are higher than nightside temperatures, but dust concentrations are higher and dust is more widespread on the nightside.

The growth phase of the B storm lasts ~ 20° of L_s from L_s = 247°–267° and is defined by the southward propagation of a warm, dusty column of air originating in the southern midlatitudes and peaking over the south pole. At the peak of the B storm (L_s = 267°), the column of dust produces maximum 50 Pa dust mixing ratios of ~ 6–8 ppm and corresponding 50 Pa temperatures around ~ 220 K. During the cessation phase, the sedimentation of dust over the south pole begins immediately following the peak of the B storm and half the dust falls out of the plume by L_s = 277°. The rate of fallout slows between L_s = 277°–287° so that half the dust present at L_s = 277° still remains at L_s = 287°.

The MGCM indicates that the 50 Pa dust in the simulated B storm is self-lofted by a pluming mechanism that cycles at an irregular interval over the course of a few sols. The plume is well defined just before solstice, developing in the eastern hemisphere at around 70° S, and it redevelops further south thereafter. The plume loses some of its structure after solstice but dust is continually lofted to 50 Pa from L_s = 267°–277°. The dust plume creates detached dust layers that are lofted 4–5 scale heights (~ 10 Pa) and then advected around the planet. Dust ascending in the plumes and the detached dust layers appear to follow the solar escalator effect. The lack of any dust lofting in the sensitivity case is especially compelling evidence that the radiative-dynamic feedbacks of airborne dust are responsible for lofting dust in the B storm.

The model-predicted dust injection rates indicate dust lifting occurs around the observed CO_2 cap edge during the B storm, especially during peak intensity (L_s = 267°), but that the MGCM-predicted surface stresses are maximized along the simulated CO_2 cap edge which is not co-located with the observed CO_2 cap edge. In the baseline
simulation, the simulated CO$_2$ cap sublimates away by $L_s = 257^\circ$ which precludes the development of surface stresses during the B storm. Raising the seasonal CO$_2$ ice albedo from 0.51 to 0.8 causes the seasonal CO$_2$ cap to persist $\sim 20^\circ$ of $L_s$ longer which better represents the retreat of the observed CO$_2$ cap and which produces surface stresses along the simulated CO$_2$ cap edge that are in better agreement with the simulated dust injection rates. This appears to indicate cap-edge dust lifting occurs along the southern CO$_2$ cap edge during the B storm.

This study provides a robust characterization of the observed B storm and contributes to the ongoing effort to better understand the behavior of dust in Martian dust storms. Expanding on the observational analysis of the B storm by Kass et al. (2016), this work presents the observed vertical structure of the B storm and proposes definitions for the growth, peak, and decay phases of the storm. The utility of GCMs as instruments that not only provide information that is lacking in observations but that also provide greater context in which to interpret observations was highlighted in the modeling analysis. The conclusions posed here regarding dust lifting and lofting in the B storm expand on previous studies that highlight the importance of cap-edge lifting in the southern hemisphere of Mars and that predict a solar escalator effect on Mars. This work underscores the importance of dust radiative heating in developing and sustaining dust storms on Mars, and similar studies could benefit from placing greater attention on these radiative-dynamic feedbacks. We are looking forward to improving the MGCM-simulated seasonal CO$_2$ cap retreat and exploring how these processes operate in the simulated B storm in the near future.
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