An Investigation of Martian Dust Devil Characteristics

A thesis submitted in partial fulfilment of

the requirements for the degree of

Doctor of Philosophy

by

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7o, My Parents

Declaration

I declare here that this thesis report represents my own ideas in my own words and I have included others ideas with appropriate citations from original sources. I also declare that I have followed all principles of academic honesty and integrity and have not misrepresented or fabricated or falsified any idea/fact/source/data in my submission. I understand that any violation of the above can cause disciplinary action by the Institute and can also evoke penal action from the sources which have thus not been properly cited or from whom proper permission has not been taken when needed.

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CERTIFICATE

It is certified that the work contained in the thesis titled **"An Investigation of Martian Dust Devil Characteristics"** by **Ms. Shefali Uttam** (Roll no: 15330010), has been carried out under my supervision and that this work has not been submitted elsewhere for a degree.

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Acknowledgments

My journey of Ph.D. at Physical Research Laboratory (PRL), Ahmedabad, which started in 2015, has now come to an end; and finally, I am ready with my Ph.D. thesis. During this journey of my Ph.D. I learned that it is not just about getting a degree, but it is to learn to believe in oneself and never lose hope. Writing this particular section of my thesis gives me immense pleasure to express my gratitude to the people who gave their support, and have directly or indirectly motivated me throughout.

First and foremost, I wish to express my deepest gratitude to my supervisor, Prof. Varun Sheel, for providing continuous support during my Ph.D. I thank him for giving me the freedom to work on new ideas and encouraging me to learn new things. His positive attitude towards solving scientific problems encourages me to probe deeper into the subject. His valuable suggestions and critical comments have helped me to improve scientific outcomes. His constant guidance, inspiration, encouragement, and motivation during this entire period helped me at the time of research and writing of this thesis.

Further, I express my sincere gratitude towards my Doctoral Scientific Committee (DSC) members, Prof. S. A. Haider and Dr. L. K. Sahu, who have always guided me during my thesis progress and provided constructive remarks to my research work. I am thankful to both of them for providing me with an unending support and their friendly behavior, which always kept me motivated for my thesis work. I am also thankful to the Academic Committee members for their insightful comments and encouragement during my research period. I would also like to thank Dr. Sanjay K. Mishra for always being available for any scientific discussion. I express my sincere thanks to all the faculty and staff members of the Planetary Sciences Division for their assistance in the form of academic and administrative guidance.

I am very grateful to the Director, Prof. Anil Bhardwaj; Dean, Prof. D. Pallamraju; and Head of Academic Services, Dr. Bhushit Vaishnav for their constant support and assistance during my Ph.D. tenure to complete the thesis work. I thank the people from accounts, purchase, library, administration, canteen, CMD, dispensary, transport, and housekeeping sections for all the help they have provided me during my Ph.D. tenure. I am also thankful to the Computer Center staff, namely Jigar sir, Tejas sir, Mishra ji, Vaibhav and others for their extended help regarding computational problems and the use of the High Performance Computation (HPC) facility. I deeply appreciate their support for providing me a smooth computational facility at all times.

I thank all the faculties of PRL for providing an overall view of research works carried out in PRL, which helped me to have a brief knowledge about different fields of sciences. I am grateful to Dr. Kinsuk Acharyya, Dr. G. S. Samanta, Dr. Arvind Singh, Pallavi Singh, and Dr. L. K. Sahu for their encouraging and friendly behavior, which always motivated me during my Ph.D. journey and eased me at various stages of my thesis work. I would also like to extend my special regards to all my teachers who taught me at different stages of my professional career. Because of their teaching, it is possible for me to reach a stage where I could write this thesis. The inspiration, support, and cooperation that I have received from my group members are beyond the scope of any acknowledgement. I would still like to thank all my group members for helping me during my Ph.D. journey. At first, I would like to thank Dr. Deepak Singh for patiently listening to all my queries and helping me find solutions for them. He is the person I approach first whenever I have any issues related to MATLAB. I would also like to thank Dr. Ashimananda Modak and Dr. Siddhi Shah for their continuous support and care in both personal and professional fronts throughout my journey. I would especially like to thank Dr. Disha Sawant, who was like my family far from home. Her wonderful suggestions and immense support kept me going through this journey. Last but not least, I would like to express my heartfelt gratitude to Masoom Jethwa, Dr. Tikemani Bag, Dr. Priyabrata Das, Dr. P. Thirupathaiah, Alka, Sonam Jiterwal, Srirag Nambiar, Dinesh Radhakrishnan, and Rashmi for their moral support and for creating a funloving working atmosphere.

Doing a Ph.D. is an extremely challenging work, which occasionally brings stress with it. It is this time when the help and support of friends relieve the stress and makes the journey memorable. I want to extend my gratitude to all my friends for making my stay in PRL comfortable. It would be impossible for me to deny the unwavering support and love that I have received from Nidhi in these five long years of my Ph.D. journey. I would like to especially thank her for all the care which she provided me with and for being the first person to approach, whenever I am in need. I would also like to thank my batch-mates; Varun, Kaustav, Archita, Subir, Aarthy, Ranadeep, Rahul, Shivangi, Akanksha, and Richa for the stressrelieving and funny moments that we shared together. Their instant help and positive support at all times kept me going. A special mention goes to Arun bhaiya for his academic guidance and being a wonderful host at his quarters. It was a pleasure to know someone like him, who is so selfless. I would also like to thank Abdur and Priyank for all the wonderful time and food which we shared at the hostel. Our deep "important" discussion on anything and everything always lighted up my mood.

Words would fail in expressing the love, strength, and encouragement of my family members towards me. I am lifelong grateful to my beloved parents, Smt. Indra Shaw and Shri Uttam Shaw, who always believed in me and encouraged me to pursue my dreams. They are my pillars of strength and have provided me with all kinds of support whenever I needed it. A special thanks goes to Ayan for being the best support anyone can get. His continuous encouragement and care helped me to go through any tough situation easily. My thesis will be incomplete if I don't express my love and thanks to all my family members; Nanaji, Nani, Sangita mausi, Rajesh mausa, Dolly mausi, Manoj mausa, Babli mausi, Soni mausi, Prabhat mausa, Dadi, Gautam uncle, Deepa aunty, Lala uncle, Upma aunty, Lappa uncle, Sapna aunty, Pramod uncle, Anita aunty, and all my cousins, who were always watching my back. Without their support, it would not be easy for me to reach this point in my life. In the end and above all, I would like to thank God.

Shefali Uttam

Abstract

The Martian Planetary Boundary Layer (PBL) is the lowest 10 km of the atmosphere that is directly influenced by the surface forcing, leading to a strong vertical mixing in the atmosphere. The dust, entrained into the PBL and consequently transported to large distances, affects the thermal and dynamical state of the Martian atmosphere. Thus, it is important to study the processes by which dust can enter in to the atmosphere, for which convective vortices have been proposed as an efficient mechanism. Convective vortices form due to heating of the surface by solar flux, leading to the formation of unstably stratified atmosphere, and hence a strong convection. The winds within the vortices have vorticity in it which makes the structure rotate. They are also accompanied by a pressure drop inside the vortex. The vortices in which winds and pressure drop are strong enough to pull dust from the surface into the atmosphere are called "dust devils".

The objective of my thesis is to study various characteristics of Martian dust devils, and their quantitative investigation. The research work comprising my thesis is based on the steady state vortex systems. Mean physical conditions, such as that for vortex wind speeds, size of vortex, atmospheric pressure and density, have been used to study vortex properties and characteristics using analytical and numerical approaches. Characteristics of convective vortices and dust devils have also been studied based on observations from a rover on Mars.

Using an analytic approach to solve the Navier Stokes equation, we have derived a simple equation for the mean tangential velocity for a vortex. Unlike other analytical solutions for vortices, our solution is dependent on the few atmospheric parameters that vary with altitude. Later, we use in-situ meteorological data recorded by NASA's Curiosity rover, to detect convective vortex events on Mars. Some of these vortex events show obscuration in solar flux data which are identified as a possible dust devils. A seasonal variation of these events suggest that they are frequent during the local summer season, during which the pressure drop associated with the vortex events are also high compared to other seasons. Lifting and distribution of dust are yet not well estimated within dust devils, for which we have provided a quantitative estimate, based on our studies. Our simulations indicate a maximum concentration of particles near the surface and at the boundary of the vortex. The larger sized particles are lifted to lower heights, as compared to its smaller counterparts. This supports the fact that within a dust devil the dust particles are distributed as per their mass along the height. The larger (heavier) particles are mostly present in lower part of the dust devils whereas, the smaller (lighter) particles move to higher heights. Finally, we have modeled electric fields within the dust devils. Due to tribocharging the dust particles get charged depending on their size. A charge separation within the dust devil gives rise to an electric field, which has important potential in lightning generation. We find that the electric field reaches the breakdown value faster in dust storm scenario as compared to non-dust storm scenario. A vertical exponential distribution of charged particles will lead to development of lower electric field than the dipolar configuration of charged particles.

Keywords: Mars; Martian atmosphere; Dust Devil; Boundary layer; Convective Vortex; Dust lifting; Electric field

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Abbreviations

AU	Astronomical Unit
CTX	Context Camera
DDA	Dust Devil Activity
DDT	Dust Devil Tracks
DOM	digital optical method
ESA	European Space Agency
EUV	Extreme Ultraviolet
FWHM	Full Width at Half Maximum
GCM	General Circulation Model
HiRISE	High Resolution Imaging Experiment
HRSC	High Resolution Stereo Camera
ISRO	Indian Space Research Organisation
LES	Large Eddy Simulations
LMST	Local Mean Solar Time
LOS	Line of Sight
Ls	Solar Longitude
LTST	Local True Solar Time
MARCI	Mars Color Imager
MCC	Mars Color Camera
MCD	Mars Climate Database
MCS	Mars Climate Sounder
MER	Mars Exploration Rover
MEX	Mars Express
MGS	Mars Global Surveyor

MOC	Mars Orbiter Camera
MOLA	Mars Orbiter Laser Altimeter
МОМ	Mars Orbiter Mission
MPF	Mars Pathfinder Lander
MRO	Mars Reconnaissance Orbiter
MSL	Mars Science Laboratory
MY	Martian Year
NCAR	National Center for Atmospheric Research
NS	Navier-Stokes Equation
PBL	Planetary Boundary Layer
PRL	Physical Research Laboratory
REMS	Rover Environmental Monitoring Station
TES	Thermal Emission Spectrometer
THEMIS	Thermal Emission Imaging System
UTC	Coordinated Universal Time
UV	Ultraviolet
WRF	Weather Research and Forecasting

Chapter 1 Introduction

Mars is roughly half the size of Earth and is farther from the Sun as compared to Earth, but it shows a striking similarity to Earth. Like Earth, Mars has an atmosphere (winds, clouds), volcanoes, and valleys. However, its atmosphere being thinner than that of Earth's, Mars is frequented by more dust storms. Even though at the present time Mars has a thinner atmosphere than Earth and no water bodies to support life, but the existence of polar ice caps and dry river channels suggests that it could have been habitable in the past [1]. More than 3.8 billion years ago, Mars was warm and wet [2] with higher average surface temperature and pressure. Studies dealing with the mapping of Mars crust suggests that early Mars had a global magnetic field which disappeared later during the evolution [3]. Since Mars experienced a loss in the intrinsic magnetic field, the energetic particles in the solar wind entered the Martian atmosphere and ionized the atmospheric constituents [4, 5]. This effect, along with low gravitational pull and photochemistry of the Mars, led to the wipe-out of the atmosphere [6, 7]. Thus, Mars acts as a natural laboratory right next to Earth, and studying about it can help scientists to understand our planet Earth more and also find a new habitable zone for humans in the near future.

1.1 Planetary features of Mars

Mars is approximately 1.52 AU distance apart from the Sun as compared to 1 AU distance of Earth from the Sun. This leads to the energy per unit area of solar radiation reaching Mars (mean solar constant, F_0) to be roughly half of that reaching the Earth. Mars being ten times lighter than Earth, possesses almost three times smaller gravitational pull. Due to this low gravity Mars cannot hold onto its atmosphere, thus making the Martian atmosphere thinner as compared to Earth. The radius of Mars is nearly half the radius of Earth, and it rotates about its axis, which is at an inclination of $\sim 25.2^{\circ}$ (obliquity), similar to that of Earth. Since the obliquity of Mars and Earth are the same, this would imply that the tropical regions of Mars are heated more intensely by the Sun than the Polar Regions. This will also imply that Mars will experience strong seasonal variations at mid and high latitudes, with each hemisphere alternating from summer to winter season in a year. Therefore, just as on Earth, there are seasons on Mars as well. The rotation frequency of is ~0.97 times smaller than Earth, hence a typical day on Mars is 24 hours, 39 minutes and 35 seconds long as compared to 24 hour day on Earth. Each Martian day is termed as Sol. Since Mars is approximately 1.5 times more distant from the Sun as compared to Earth, it takes more time than Earth to complete one full revolution around the Sun. One Martian year is 668.6 sols (~687 Earth days) long, approximately double to that of Earth's (365.25 days). Table 1.1 lists the key planetary parameters for both Mars and Earth.

Table 1.1: The key planetary parameters for both Mars and Earth [Haberle et al.,2017 [8]; Read and Lewis, 2004 [9]; Sanchez-Lavega, 2011 [10]].

Planetary Parameters	Mars	Earth	Ratio
Radius (km)	3396	6378	0.53
Mass (kg)	6.39×10 ²³	5.97×10 ²⁴	0.11
Average distance from Sun (AU)	1.52	1	1.52
Mean solar constant (Wm ⁻²)	589	1367	0.43
Orbital eccentricity	0.093	0.017	5.47
Length of year (Mars sols)	668.6	355.6	1.88
Length of year (Earth days)	687.0	365.25	1.88
Spin-axis inclination (°)	25.19	23.44	1.08
Rotation frequency (s ⁻¹)	7.09×10 ⁻⁵	7.29×10 ⁻⁵	0.97
Solar day (h)	24.66	24.00	1.03
Surface gravity (ms ⁻²)	3.71	9.80	0.38
Average surface pressure (hPa)	6.1	1013	0.006
Average surface temperature (K)	210	290	0.72
Average atmospheric density at	0.02	1.2	0.017
surface (kgm ⁻³)			
Gas constant (JK ⁻¹ kg ⁻¹)	191	287	0.66

Mars revolves around the Sun in an elliptical orbit having an orbital eccentricity of 0.093, whereas the orbit of Earth is relatively circular. An elliptical orbit leads to an asymmetry in duration of the seasons between the northern and southern hemispheres. Table 1.2 lists the duration of the season on Mars and Earth in terms of Ls, number of sols, and days. The seasons on Mars are denoted by "Solar Longitude". The Solar Longitude (Ls) or Aerocentric Longitude is the angle between the Sun and Mars along the Mars orbit, which is measured from northern hemisphere spring equinox (Ls = 0°). The perihelion occurs during southern summer (Ls = 251°), thus making southern summer relatively short but intense, while aphelion occurs during southern winter (Ls = 71°), thus making it relatively long and mild. Figure 1.1 is a schematic representation of the seasons on Mars along its path of revolution around the Sun in its elliptical orbit.



Figure 1.1: A schematic representation of the seasons on Mars.

The timekeeping on Mars started from April 11, 1955, which is from the beginning of vernal equinox on Mars ($Ls = 0^{\circ}$) [11]. It was this time when the systematic scientific observation and documentation of data started for the Martian atmosphere. At present we are in Martian Year (MY) 35, which extends from March 23, 2019 to February 7, 2021.

Season	Length of Season on	Length of Season on
(Northern Hemisphere)	Mars	Earth
	(sols)	(days)
Spring (Ls = 0° to 90°)	193.3	92.8
Summer (Ls = 90° to 180°)	178.6	93.6
Autumn (Ls = 180° to	142.7	89.8
270°)		
Winter (Ls = 270° to 360°)	154.0	89.0

Table 1.2: The duration of the season on Mars and Earth in terms of Ls, number of sols, and days.

1.2 Atmospheric composition of Mars

The atmospheric number density on the surface is $\sim 10^{17}$ cm⁻³ for Mars as compared to $\sim 10^{19}$ cm⁻³ for Earth, which makes Martian atmosphere approximately 100

times thinner than that of Earth. The gas composition of Mars is primarily studied using both ground-based (rovers and landers) and spacecraft-based spectroscopic instruments. Its atmosphere comprises mostly of carbon dioxide (~95%) with small amounts of nitrogen (~2%) and argon (~2%). Oxygen, carbon monoxide, ozone, and water vapor are the trace constituent of the atmosphere, with concentrations present in variable amounts (variable in space and time) ranging from parts per thousands to parts per billion levels. Table 1.3 lists the atmospheric composition for both Mars and Earth's atmosphere [12–17].

Table 1.3: The atmospheric composition for both Mars and Earth's atmosphere [Clancy et al., 2016 [12]; Franz et al., 2015 [13]; Lagzi et al., 2014 [14]; Perrier et al., 2006 [15]; Smith 2004 [16]; Trainer et al., 2019 [17]].

Atmospheric Species	Concentration on Mars	Concentration on Earth
Carbon dioxide CO ₂	95.1 %	0.039 %
Nitrogen N ₂	2.59 %	78.084 %
Argon Ar	1.94 %	0.934 %
Oxygen O ₂	0.161 %	20.946 %
Carbon monoxide CO	0.058 %	0.05 – 0.25 ppm
Water vapor H ₂ O	15 – 1500 ppm	0-4 %
Ozone O ₃	10 – 350 ppb	0.01 – 0.05 ppm

The Martian atmosphere is relatively dry as compared to Earth's atmosphere, due to less water vapor concentration. Even though the water vapor is sparse in the atmosphere, both water vapor and carbon dioxide condense to form thin ice clouds [9]. Due to the condensation-sublimation cycle, the global abundance of carbon dioxide varies by ~30 % on a seasonal basis through condensation and sublimation from seasonal ice caps at each pole [8, 18]. This leads to a significant seasonal variation in the concentration of carbon dioxide and other non-condensable species (e.g., argon, oxygen, carbon monoxide) in the atmosphere. Apart from these gaseous species, Martian atmosphere is also laden with fine dust particles at all time. The concentration of dust in the atmosphere varies with season and year. More details about Martian dust will be discussed in section 1.5.

1.3 Pressure, Density and temperature variations on Mars

The surface pressure gives a direct indication of the column-integrated mass $\left(\frac{P_o}{g}\right)$ of the atmosphere. Since the major constituent of the Martian atmosphere is CO₂, the surface pressure will roughly correspond to its column abundance. The first estimation of surface pressure was done using ground-based spectroscopy of CO₂ lines [19] and radio occultation from the orbit by Mariner 4 [20] with reasonable accuracy. The first direct measurement of Mars atmosphere from Mariner 4 shows the surface pressure to be varying between 4 to 7 mbar [20]. Later several landers, rovers and orbiting spacecraft provided accurate measurements of surface pressure at different

geographical regions on Mars [21–26] indicating an average surface pressure of Martian atmosphere as 6.1 mbar. The Martian surface pressure is roughly 100 times smaller than that of Earth's, and the hydrostatic law $(P(z) = P_0 e^{-(Z/H)})$ gives its variation in the vertical direction. According to the hydrostatic law, surface pressure decreases exponentially with increasing altitude in accordance with a length scale known as the "scale height (H)". The scale height of Mars is typically about 10 km [9], which means that an increase in the altitude by this magnitude will lead to a decrease in surface pressure by a factor of *e*. The scale height for Earth is approximately 8 km [9] due to which the rate of pressure decrease in vertical direction is higher for Earth as compared to Mars. Figure 1.2 shows the seasonal variation of surface pressure at the location of Viking Lander 1 (22.27°N, 312.05°E) and Viking Lander 2 (47.64°N, 225.71°W) during the first Martian year of Viking's operation [27–30].



Figure 1.2: Seasonal variation of surface pressure at the location of Viking Lander 1 (22.27°N, 312.05°E) and Viking Lander 2 (47.64°N, 225.71°W) during the first Martian year of Viking's operation [Image credit: Hess et al., 1976a [27], 1976b [28]].

The offset between the two pressure curves in figure 1.2 is primarily caused by the elevation difference (~1.2 km) between the two landing sites. These in-situ measurements show that the surface pressure varies by roughly 30%, and it also undergoes a seasonal cycle [18, 29]. This phenomenon occurs due to consecutive condensation and sublimation of CO₂. At the winter pole, CO₂ condenses to form a seasonal ice cap, thus reducing its columnar concentration in the atmosphere and thereby reducing the surface pressure. In summer, polar CO₂ sublimates and leads to an increase in both CO₂ columnar concentration in the atmosphere and surface pressure. The air density also varies exponentially in the vertical direction for the Martian atmosphere following the hydrostatic law. An important constituent of any atmosphere is the water vapor concentration, since it is a greenhouse gas and leads to heating of the atmosphere. As already described in section 1.2, the water vapor is a trace gas in the Martian atmosphere whose abundance varies greatly with season and location. The atmospheric models show that the vertical distribution of water vapor in Martian atmosphere is exponential [31]. But recent observations of vertical distribution of water vapor shows that there is a presence of detached layers enriched in water [32]. The hygropause (altitude at which water vapor saturation occurs) on Mars varies with season and location and is reported to be at ~40 km altitude with the mixing ratio in the order of 10^{-4} [33]. Above this height, the concentration of water vapor decreases significantly. On Earth, the detached layer of water vapor is not present and its concentration decreases rapidly as we move higher in altitude. Approximately 99% of water vapor is constratined in the troposphere on Earth [34]. Hence, at ~40 km we do not expect any significant amout of water vapor to be present in the Earth's atmosphere.

Vertical regions in the Martian atmosphere are defined by analogy to those on Earth but with some differences. Based on an average temperature profile, the Martian atmosphere is divided into four layers: the troposphere, mesosphere, thermosphere and exosphere. An average global profile of temperature, derived from the Martian Climate Database (MCD) [35] is shown in figure 1.3 [36]. The shaded region in the figure represents the seasonal extent of the vertical temperature profile.



Figure 1.3: An average temperature profile of the Martian atmosphere derived from the Mars Climate Database (MCD). The shaded region is the seasonal spread of the global averaged temperature [Image credit: Modak, 2019 [36]].

The troposphere (surface to ~50 km) is defined as that region in the atmosphere where temperatures typically decrease with altitude (z). The rate at which temperature decreases with altitude in the troposphere is defined as the lapse rate $\left(-\frac{dT}{dz}\right)$. The dry adiabatic lapse rate for Mars is ~5 K/km [8, 9], whereas, for Earth is ~9.8 K/km., which
can be accounted by the thin atmosphere of Mars. Unlike Earth, Mars has no persistent ozone layer, thus leading to an absence of a stratosphere. Instead, Mars has a mesosphere (~50 km to ~100 km) just above the troposphere, which is defined based on the vertical temperature profile. The lapse rate is lower here as compared to the lapse rate in the troposphere, due to which the decrease in temperature is less with increasing altitude. Above 100 km is the thermosphere (~100 km to ~200 km), where temperature increases with altitude due to the absorption of solar EUV (Extreme Ultraviolet) radiation. Above ~200 km, the atmospheric density becomes almost negligible, leading to an insignificant change in the temperature of the region. This region is termed as exosphere (~200 km and above). The temperature in this region is almost constant with increasing altitude and atmospheric species escape depending on their energy.

Depending on the season, time of the sol, and latitude, the atmospheric temperature near the surface can vary from a minimum of ~130 K over the northern polar region during northern winter to ~300 K over the southern tropical region during the perihelion season [37]. Figure 1.4 shows the longitude averaged global seasonal temperature variation in the Martian atmosphere close to the surface as observed by Thermal Emission Spectrometer (TES) [16].



Figure 1.4: Longitude averaged global seasonal temperature variation in the Martian atmosphere close to the surface as observed by TES [Image credit: Smith, 2004 [16]].

The seasonal temperature variation shows that as the southern hemisphere spring starts, the temperature starts to increase. The temperature in the southern hemisphere during summer is highest compared to the summer of the northern hemisphere. It is so because Mars is closest to the Sun (perihelion) during the southern summer season.

1.4 Atmospheric Boundary Layer

As already described in section 1.3, based on the temperature profile, the Martian atmosphere is divided into various layers. The troposphere extends up to 50 km in which temperature decreases monotonically. The Planetary Boundary Layer (PBL) is the lowest part of the troposphere whose behavior is directly influenced by the presence of the planetary surface. The PBL responds to surface forcings, i.e., atmosphere-surface exchanges of momentum, energy, and mass, in a timescale of an hour or less [38]. This makes the PBL highly turbulent [39, 40], and therefore, the vertical mixing is strong. The boundary layer thickness varies with space and time, depending on the turbulence in the atmosphere. On Mars, the atmospheric boundary

layer can extend up to an altitude of about 10 km as compared to about 1 - 2 km on Earth [40, 41]. Figure 1.5 shows the typical vertical structure of various layers within the Martian troposphere.



Figure 1.5: Typical vertical structure of the Martian troposphere.

The troposphere is divided into two broad zones viz. planetary boundary layer and free atmosphere (Figure 1.5 (a)). The boundary layer can be further divided into the surface layer, mixing layer, and entrainment zone (Figure 1.5 (b)). The surface layer consists of a roughness layer and a logarithmic layer (Figure 1.5 (c)). In the surface layer, the atmospheric quantities, like wind speed and temperature, vary rapidly with altitude. The turbulence in this layer is generated due to wind shear and buoyancy effects caused by the heating of the surface. The logarithmic layer is so called because the vertical profile of horizontal wind velocity is approximately logarithmic in this layer. The expression for the mean horizontal wind velocity in the Cartesian co-ordinate system is given as [42]:

$$\overline{u} = \frac{u_*}{k} \ln \left(\frac{z}{z_o} \right) \qquad \dots (1.1)$$

where, u_* is the friction velocity (often termed as drag velocity) near the surface, 'k' is von Karman constant, z is the altitude, and z_0 is the aerodynamic roughness length of the surface. The friction velocity is not the actual wind velocity but a scaling parameter for the force exerted by wind on the surface [38]. The roughness parameter measures the effectiveness of a surface structure to absorb momentum and is a function of factors like topography (scale of meters) and the size of the dust particles on the surface [43, 44]. It can also be defined as the height where the extrapolated wind flow approaches zero or at which the turbulent fluid flow becomes laminar. The above semi-empirical logarithmic variation of horizontal wind velocity is not defined in the mixing layer, due to high fluctuations in physical parameters like pressure, temperature, and density. The Earth's daytime surface layer extends up to about 0.1 km [45]. However, the Mars' daytime surface layer can extend up to 1 km [40]. The surface layer of PBL is always turbulent, and the major contributor to this turbulence is the daytime convection. Therefore, the PBL is also sometimes called the daytime convective layer. This can lead to formation of small-scale features such as convective vortices in the PBL, which are called dust devils if they are strong enough to lift dust [46–49]. Heating of the surface by solar flux under certain conditions leads to a super adiabatic lapse rate (lapse rate greater than the dry-adiabatic lapse rate of the atmosphere) in the lower part of PBL. This causes the formation of an unstably stratified atmosphere, leading to a strong convection [50, 51], under which vortices can form.

1.5 Martian dust

Dust is an important component of the Martian climate system. Gierasch and Goody (1972) [52] were the first to recognize the influence of the radiatively active airborne dust on the thermal and dynamical state of the Martian atmosphere. The dust absorbs and scatters solar radiation at visible, ultraviolet [15] and infrared wavelengths [52]. The absorption of infrared radiation by dust results in the heating of the atmosphere. The evidence for the presence of dust on Mars was first given in the 18th century by telescopic observation of the yellow clouds in the Martian atmosphere [53, 54]. The Mariner 9 orbiter in 1971 observed the Martian atmosphere to be completely enveloped by airborne dust [55, 56]. Later in-situ observation of the atmosphere by Viking landers also confirmed the presence of dust in the atmosphere of Mars [57]. The effect of dust, entrained into the atmosphere from the surface, on the atmosphere and climate have been studied using observations and numerical models [52, 57–63]. The long term observation of Martian atmosphere by Thermal Emission Imaging System (THEMIS) onboard Mars Odyssey mission and Thermal Emission Spectrometer (TES) onboard Mars Global Surveyor (MGS) provides an insight about the seasonal and yearly trends of dust cycle on Mars. Observations indicate that dust remains suspended in the Martian atmosphere for the whole year with seasonally and yearly varying abundances, i.e., the dust number density at surface during non-dust storm year is ~ 3 cm⁻³ whereas, the dust number density at surface during dust storm year can reach up

to ~13 cm⁻³ [64]. The dust cycle on Mars, Figure 1.6 shows the seasonal variability of zonally averaged dust optical depth at 9.3 μ m band retrieved from the THEMIS onboard Mars Odyssey mission [65, 66].



Figure 1.6: Zonally averaged dust optical depth at 9.3 µm band as a function of Ls and latitude, retrieved from THEMIS for more than 3.5 Martian years. [Image credit: Smith, 2009 [65]]

Dust can be lifted from the surface into the atmosphere by various processes such as surface wind stress and dust devils [46] and will be discussed in detail later in this chapter. These dust lifting processes entrain dust in the atmosphere which can remain suspended for a long time as the gravitational pull of Mars is low. This dust can then be transported by winds to spatial scales from local to global, causing the dust to remain suspended from less than an hour to seasonal timescales [8, 9]. In this section, we will discuss the physical properties of dust and methods by which it is injected into the atmosphere.

1.5.1 Dust properties

Based on the infrared spectra observed by orbiting spacecrafts [67–69], it is believed that the Martian soil has silicate mineral (SiO_2) as the dominant component (~50 wt% [70]). Mars is often described as the "Red Planet" due to the presence of a high amount of iron oxides in the Martian soil [71]. The composition, size, and shape of the dust particles in the Martian atmosphere, determine their optical properties. As already described in earlier sections that dust remains suspended in the atmosphere, it affects the scattering of solar radiation. Since the size of dust particles is larger than the air molecules, it becomes comparable to the wavelength leading to Mie scattering in the atmosphere. Due to the Mie scattering, the sky on Mars appears pinkish-red [72– 74]. The properties of Martian dust are inferred indirectly from various optical measurements of extinction, scattering intensity, and polarization over ranges of wavelengths. These measurements are then compared with the predictions of Mie theory [75], which allows the computation of integrated optical properties, like refractive index, single scattering albedo, and extinction efficiency; of homogeneously dispersed, uniform particles of spherical shapes [76]. The particle size distribution of dust aerosols is dependent on the first and second moments of the surface-areaweighted particle size distribution: the effective radius (r_{eff}), and the effective variance (v_{eff}) respectively [77]. A modified gamma distribution [64, 78] is assumed for the dust particle size distribution in the Martian atmosphere, which is expressed as,

$$n(r) = r^{a} \exp\left(-br\right) \qquad \dots (1.2)$$

where, *r* is the radius of the spherical-shaped dust particle, n(r) is the number of dust particles having radius between *r* and r + dr, $a = \frac{1-3v_{eff}}{v_{eff}}$ and $b = \frac{1}{v_{eff}.r_{eff}}$. During dusty scenario, the effective radius defined above, is higher as compared to conditions of lower dust loading, at a given altitude [79–82]. In general, one can consider the effective radius for Martian dust particles varying between 1.2 to 1.8 µm and the effective variance for Martian dust particles varying between 0.2 to 0.4 [76, 78]. Typically, the particle density of the material constituting Martian dust is approximately 2600 kg.m⁻³ [83].

The vertical dust distribution greatly influences the temperature structure of the atmosphere and also affects the surface temperature [57]. Analysis of Mariner 9 orbiter retrieved dust opacity led to the widely used "Conrath" profile describing the typical vertical dust distribution [84]. The Conrath dust profile is a profile for representing dust distributions in the Martian atmosphere, derived by balancing upward transport of dust due to diffusive mixing and downward gravitational sedimentation. The Conrath profile shows that the maximum dust is near the surface, and it decreases as we move higher in altitude. The profile is expressed as,

$$q(z) = q_o \exp\left[\nu\left(1 - e^{-(z_H)}\right)\right] \qquad \dots (1.3)$$

where, q(z) is dust mixing ratio at altitude z, q_o is dust mixing ratio at the reference altitude (mostly surface), v is the Conrath parameter, which determines how deeply the dust is mixed and H is the scale height of the atmosphere. This profile was widely used in General Circulation Models (GCM) to investigate aspects of the dust cycle and its effect on the climate of Mars [85–87]. However, in recent years limb observations of Thermal Emission Spectrometer (TES), onboard Mars Global Surveyor (MGS), and Mars Climate Sounder (MCS), onboard Mars Reconnaissance Orbiter (MRO), have revealed the existence of local maximum of dust mixing ratio at higher altitudes and not the surface, at particular locations and seasons [88–92]. These elevated dust layers are also termed as "detached dust layers" and are observed throughout much of the year at tropical and subtropical latitudes [64, 89-92]. These detached dust layers could have significant implications on the thermal structure of the atmosphere and are beginning to be explored by incorporating MCS or TES observed profiles into GCMs. Due to unresolved vertical transport processes in GCMs, they only predict the presence of detached dust layer during southern spring and summer but cannot predict the same during northern spring and summer [49, 93]. The Mesoscale modeling of dust transport can produce detached dust layer by associating local and regional scale phenomena with topographic flows and radiative-dynamic feedbacks between dust heating and their circulation [94–97]. Hence, to understand the dust loading in the atmosphere, we must understand the methods which entrain or remove dust from the atmosphere. We will discuss this in the next two sections, where the dust lifting into the atmosphere and its deposition back on the surface is described.

1.5.2 Dust lifting theories

The processes by which dust particles can be raised from the surface of Mars and enter the atmosphere has been an important topic for research. The study of dust lifting on Earth dates back to studies related to sand dunes, but for Mars the importance of dust loading in the atmosphere on Martian climate led to renewed interest in the theory of dust lifting mechanisms [98–101]. Here we will discuss two processes that are believed to be responsible for dust lifting from the surface of Mars. The first is dust lifting due to surface wind stress, which includes direct injection of dust particles and dust lifting via saltation. The second process is lifting by convective vortices or "dust devils", in which strong vertical motion of convection-driven vortex winds is responsible for dust injection into the atmosphere.

1.5.2.1 Lifting due to surface wind stress

Based on the observations of wind-blown sand and dust in the North African desert, Robert Bagnold [102] suggested that during the lifting process, the surface particles provide resistance to the blowing surface winds. This resistance is termed as the fluid threshold, which is defined as the speed at which wind stress can alone enable lifting of dust particles directly from the surface [9]. This fluid threshold is represented by threshold drag velocity (u_*^{th}) , which must be exceeded by the drag velocity (u_*) of the blowing wind in order for the lifting of dust to occur. This drag velocity is dependent on the near-surface wind stress (ζ) and atmospheric density (ρ) ,

$$u_* = \sqrt{\frac{\zeta}{\rho}} \qquad \dots (1.4)$$

Within the lower part of the atmosphere, the horizontal wind velocities vary approximately logarithmically with height, which has already been defined in equation (1.1). This wind exerts forces on the surface dust particle, as shown in figure 1.7 [45].



Figure 1.7: Forces acting on a particle lying on the surface under the influence of blowing wind [Image credit: Shao, 2008 [45]].

The forces acting on the dust particles in the presence of a wind field are: drag force (F_d) , lift force (F_l) , gravity force (F_g) and inter-particle cohesive force (F_i) . At the instant when dust particle starts its motion, the combined effect responsible for lifting the particle $(F_d \text{ and } F_l)$ will be balanced by the combined retarding effect $(F_g \text{ and } F_i)$. The balance of forces at the instant of particle lift-off can be obtained by the summation of moments about the point P as,

$$r_d F_d + r_i F_i \approx r_i F_a + r_i F_i \qquad \dots (1.5)$$

where, r_d , r_l and r_i are moment arm lengths. A simple determination of the threshold drag velocity (u_*^{th}) for a spherical loose particle on a smooth surface, is by considering only the balance between the aerodynamic drag and the gravity force (using equation (1.5)) [102, 103],

$$u_*^{th} = A_{\sqrt{\frac{\left(\rho_p - \rho\right)g D_p}{\rho}}} \dots (1.6)$$

where, ρ_p is dust particle density, ρ is the atmospheric air density, g is the acceleration due to gravity, D_p is the mean particle diameter, and A is the semi-empirical expression for the threshold parameter obtained by Greeley and Iversen (1985) [104] in wind tunnel experiments. This parameter A is explicitly dependent on the inter-particle cohesion (I_p) between particles, particle size, and the friction Reynolds number at the threshold. From numerical solutions of the relations determining A [46, 104, 105], the final dependence of u_*^{th} on particle size and inter-particle cohesion can be determined and is shown in figure 1.8 for the Martian atmosphere [46].



Figure 1.8: Variation of threshold drag velocity (left axis) and corresponding wind speed at 5m above the surface (right axis) with particle diameter and different interparticle cohesion (I_p) [Image credit: Newman et al., 2002a [46]].

We can see from figure 1.8 that although smaller particles are lighter than larger particles for a given particle density, they experience a stronger inter-particle cohesion effect than larger particles, as cohesion increases with the ratio of surface area to volume. Due to this, smaller particles become harder to lift. Hence, there is an optimal particle size which can be lifted from the surface for a given value of I_p , like particles of diameter ranging between 50 µm to 150 µm diameter (sand-sized) are most easily lifted for I_p between 10^{-7} and 10^{-6} Nm^{-1/2}. Since it is harder to lift smaller sized particles by wind stress alone, the mechanism of saltation has been suggested. Saltation is a method when larger particles (already lifted from the surface) fall back on the surface and help to eject smaller particles from the surface [106–109]. Bagnold (1954) [102] and White (1979) [110] found that the upward flux of small particles lifted, once the fluid threshold for large particles had been exceeded, could be related to the horizontal flux of large particles, F_H , [9, 46],

$$F_{H} = \max\left[0, \ 2.61 \frac{\rho}{g} (u_{*})^{3} \left(1 - \frac{u_{*}^{th}}{u_{*}}\right) \left(1 + \frac{u_{*}^{th}}{u_{*}}\right)^{2}\right] \qquad \dots (1.7)$$

Currently, equation (1.7) is used in GCMs to parameterize the lifting of dust in the Martian atmosphere due to wind stress. Once the dust particles are lifted into the lowest region of the atmosphere, the turbulence in the atmosphere can be expected to mix the dust upwards.

1.5.2.2 Lifting due to convective vortices

Another way by which surface dust can enter the atmosphere is by entrainment into regions of strong vertical motion generated due to thermally-driven convection in the atmosphere. Figure 1.9 shows the schematic diagram of a convection generated vortex in the atmosphere.



Figure 1.9: A schematic diagram of a convection generated vortex in the atmosphere.

A convective vortex forms due to surface heating, which under certain conditions can develop convection and lead to the generation of plumes of rising warm air. This rising warm air then interacts with cooler air at the top in such a way as to produce initial rotation due to the conservation of angular momentum [111]. Once a rotating core is formed, the column of warm air starts to rise and move vertically. This upward motion in the vortex core just above the surface produces a pressure drop, leading to more warm air being pulled inside the core, whereas the cooler air at the top starts falling downwards. This further intensifies the vortex until it is self-sustaining, with peak tangential velocities near the surface at the boundary of the core. Sometimes, the winds and pressure drop inside this vortex is strong enough to pull the dust from the surface into the atmosphere, leading to the formation of "dust devils". Active dust devils were first observed from orbit by Viking Orbiter cameras [112]. They have since been observed as a long bright column of dust casting a shadow on the surface and often leaves a trail behind [113–115]. Later, subsequent observations of dust devils were made from other orbiter missions like MGS [113, 115–119], Mars Odyssey [114, 120], Mars Express (MEX) [121–126], MRO [127] and Mars Orbiter Mission (MOM) [128]. Figure 1.10 shows an image of a large dust devil, as observed by Mars Orbiter Camera (MOC) onboard MGS on Mars [9].



Figure 1.10: An image of a large dust devil as observed by Mars Orbiter Camera (MOC) onboard MGS on Mars [Image credit: Read and Lewis, 2004 [9]].

The shadow analysis method is used to determine the dimensions of the dust devils observed by the orbiters. These observations show that dust devil ranges in size from a few meters broad and tens of meters high to over a kilometer broad and more than 6 km in height. Mars Color Camera (MCC) onboard MOM has observed several dust devils while orbiting the planet Mars since September 2014. The shadow analysis of five dust devils, as observed by MCC on November 7, 2016, shows that the height of the dust devils varies between ~0.4 km to ~2 km [128]. Observations of dust devils from orbit are limited by the fixed local times of the spacecraft's orbits. Therefore, detailed information regarding the diurnal variability of dust devil activity comes from surface observations. The surface-based missions (landers and rovers) are also used to observe multiple occurrences of dust devils. The first visual detection of dust devils on Mars from a surface instrument was accomplished by Mars Pathfinder Lander (MPF) mission [129, 130]. Later other landers/rovers, like Spirit, Opportunity, Phoenix and Curiosity observed dust devils using different methods (which will be discussed in details in chapter 3) at different locations [131–137]. Figure 1.11 shows an image of a dust devil as recorded by navigation camera onboard Spirit rover in Gusev crater [138].



Figure 1.11: An image of a dust devil as recorded by navigation camera onboard Spirit rover in Gusev crater [Image credit: Whelley and Greeley, 2008 [138]].

Dust devils and their tracks have been observed in both hemispheres in all seasons using both orbiter data and surface experiments. The peak activity of dust devils occurs during regional spring and summer in each hemisphere [115, 118, 139]. Analysis of the diurnal activity of convective vortices from various rovers and landers shows that most vortices occur during the noon-time (~1200 and ~1400 local time) [41, 133–135, 137, 140, 141].

Dust devils provide an essential mechanism for listing dust into the air in localized regions. They have an important contribution in maintaining the background dust loading in the atmosphere, even under relatively clear conditions when no dust storms are occurring nearby [47, 48, 131, 138, 142]. This contribution of dust devils as a source of dust in the atmosphere has been estimated by a simple thermo-dynamical model of a dust devil [143, 144]. In this model, a convective vortex is assumed as a heat engine performing mechanical work against frictional dissipation. The vortex is driven by heating due to warm air, which is pulled into the vortex just above the surface. The mechanical energy made available to drive dust devils near the surface in the convective heat engine (F_{av}), also known as "Dust Devil Activity" (DDA) index, is given as,

$$F_{av} = \eta F_s \qquad \dots (1.8)$$

where, F_s is the surface sensible heat flux which is approximately proportional to the temperature difference of the surface and the near-surface air and η is the thermodynamic efficiency of the convective dust devil heat engine which is defined as the fraction of input heat energy transformed into mechanical work. The thermodynamic efficiency is given approximately by (1 - b), where,

$$b = \frac{\left(p_s^{\chi+1} - p_{top}^{\chi+1}\right)}{\left(p_s - p_{top}\right)(\chi+1)p_s^{\chi}} \qquad \dots (1.9)$$

where, p_s is the ambient surface pressure, p_{top} is the ambient pressure at the top of the convective boundary layer, and χ is the ratio of specific gas constant to the specific heat capacity of the atmosphere. It can be understood from equation (1.9) that as p_{top} increases, b increases, and η decreases. Thus, if the thickness of the convective boundary layer increases, the efficiency of the dust devil heat engine will increase, leading to more lifting of dust. The vortex intensity for this model is given by the pressure drop (Δp) as [143],

$$\Delta p \approx p_s \left\{ 1 - \exp\left[\left(\frac{\gamma \eta}{\gamma \eta - 1} \right) \left(\frac{\eta_H}{\chi} \right) \right] \right\} \qquad \dots (1.10)$$

where, γ is the fraction of the total dissipation of mechanical energy consumed by friction at surface and η_H is the horizontal thermodynamic efficiency of the vortex $(\eta_H \equiv \frac{T_o - \overline{T_s}}{\overline{T_s}}), \overline{T_s}$ being the average temperature of the surface air outside the vortex and T_o being the temperature at the center of the vortex near the surface. Assuming the vortex in cyclostrophic balance (the centrifugal and radial pressure gradient forces balance), the tangential wind speed (V) around the vortex is given as [143],

$$V \approx \sqrt{R\overline{T_s}} \left\{ 1 - \exp\left[\left(\frac{\gamma\eta}{\gamma\eta - 1}\right)\left(\frac{\eta_H}{\chi}\right)\right] \right\} \qquad \dots (1.11)$$

where, R is the gas constant of the air in the atmosphere. Equation (1.11) suggests that for a vortex having larger η (taller vortex since it will have deeper boundary layer) and larger η_H (wider vortex since it will have larger core-to-boundary temperature gradient), the tangential velocity will be higher. Such a model (pressure drop and tangential velocity) seems capable of predicting where dust devils are most likely to occur, and also to give some indication of their intensity and an estimate of the average flux of dust into the atmosphere can be obtained. GCMs use both equation (1.8) and equation (1.11) to predict dust lifting due to convective vortices [46, 47, 49, 87]. The dust devil activity (equation (1.8)) of convective vortices are used to determine the dust lifting by assuming that dust lifting depends primarily on the sensible heat exchange from the surface to the atmosphere and the depth of the planetary boundary layer. It is also assumed that dust lifting occurs when the tangential wind speed exceeds the threshold tangential frictional wind speed required for dust lifting. Several laboratory experiments [83, 104, 111, 145, 146] and terrestrial field experiments [50, 139, 147, 148] are also conducted to study dust devils and determine the dust flux raised by them for Earth and Mars conditions. The dust devils inject dust in the atmosphere by a combination of two mechanisms: force analogous to near surface wind stress and pressure suction effect [149]. Laboratory studies have shown that the "suction effect" generated due to the pressure drop and high tangential velocities within a convective vortex leads to the lifting of smaller (micron sized) dust particles, which was earlier tough for the near-surface winds. The effect of lifting due to near surface wind stress within vortices are mainly determind by the drag force applied by the vortex wind on particles [149]. Once the particles are dragged from their position it becomes easy to lift them. The lift force, which is accounting for the pressure suction, then acts on these particles and lifts the particles above the ground. The individual contribution of these two mechanisms in the Martian dust cycle is studied using GCMs. Kahre et al., 2006 [49] suggests that both near surface wind stress and dust devil lifting are necessary to simulate the zonally and globally averaged spatial and temporal pattern of atmospheric dust loading.

1.5.3 Dust deposition theory

Once the dust is lifted into the atmosphere, it will tend to settle back on the planetary surface with time. The processes that govern the removal of dust from the Martian atmosphere to the surface are studied since Mariner 9 provided the first spacecraft observations of the decay phase of a global dust storm of 1971-1972 [84, 150]. Later, several Mars-orbiting spacecraft observed the decay of five additional global dust storms and many local and regional dust storms [151]. In situ rates of dust accumulation are also inferred from measurements obtained by Pathfinder, Phoenix, and the Mars Exploration Rovers [16, 152–159]. On Earth, the dust removal process is divided into two categories: wet and dry, but due to the lack of significant amount of water vapor in the Martian atmosphere, the dry deposition dominates the dust sedimentation process on Mars [160]. By dry deposition of dust, one means gravitational sedimentation. Since dust particles are denser than the other constituent of the atmosphere, it will tend to fall under gravity. Hence the sedimentation velocity of the dust particles can be found by equating the buoyancy-adjusted downward force (due to gravity) and the upward drag force (due to viscous friction between the falling particles and the atmosphere). Due to the thin atmosphere of Mars, an additional correction factor to the sedimentation velocity is required, because the size of the lifted dust particles is less than the mean free path between the molecules of atmosphere. This

factor is known as the Cunningham 'slip-flow' correction [160]. The effective sedimentation velocity (w_{sed}) for the dust particles on Mars is given as [9],

$$w_{sed} = \frac{2g\left(\rho_p - \rho\right)}{9\eta_d} \left(\frac{D_p}{2}\right)^2 \left(1 + \frac{8\zeta_f}{3D_p}\right) \qquad \dots (1.12)$$

where, η_d is the molecular viscosity of the air and ζ_f is the mean free path. For a typical dust particle on Mars (size $\sim 1 \mu m$), the sedimentation velocity will be of the order $\sim 1 - 1 - 1 - 1$ 2 mm.s^{-1} , whereas for dust particle having size ~100 µm (sand-sized) the sedimentation velocity will increase to $\sim 50 - 80$ cm.s⁻¹ [9]. In the absence of any active dust lifting or vertical motion, the dust particles of size $\sim 1 \,\mu m$ will take around 1 - 2 months to settle from a height of ~10 km, whereas, for the dust particles of size ~100 μ m (suspended at ~10 km) will take only a few hours to settle down [9]. Such timescales of dust sedimentation are reasonably consistent with the observation since local sandstorms on Mars are relatively short-lived, whereas small dust particles can remain suspended in the atmosphere for long periods, thus maintain the constant background haze. Even though Mars has a scanty amount of water vapor in its atmosphere, the Polar Regions on Mars do have a larger amount of CO_2 and water vapor, which condenses during the winter season. The presence of water vapor acts as condensation nuclei for the dust and will enhance the dust deposition process in the Polar Regions. The modeling studies and the recent observations of large water-ice cloud particles near the surface of the Phoenix Lander site also show that wet deposition processes may be important at some locations during some seasons [161–163]. Several studies also propose the possibility of dust deposition due to the presence of water vapor outside the Polar Regions to be a major mechanism to shut down large dust storms on Mars [164, 165]. Parameterizations 32

have also been done in GCMs to incorporate dust deposition occurring due to CO₂ ice formation during polar winter at high latitudes.

1.6 Objectives of the thesis

The suspended dust in the Martian atmosphere comes from its source at the surface. We have discussed various mechanisms above that help in the entrainment of dust into the atmosphere. Dust has an important feedback on the climate of Mars. For example, it affects the temperature cycle of the Martian atmosphere due to the radiative heating of the atmosphere. The net effect of a layer of dust in the atmosphere is quite complicated, and it can lead to a net warming of the atmosphere due to the absorption of solar radiation [57, 58, 85, 86, 166]. During the daytime, the dust in the atmosphere absorbs the infrared radiation, heats the atmosphere in its vicinity, and does not allow infrared radiation to reach the surface. Due to this, a net cooling of the surface during daytime can be observed. However, it affects the rate at which the surface and lower atmosphere can cool during the night. When the surface emits in the infrared region, the dust present in the atmosphere does not allow this energy to escape and leads to more warming of the lower atmosphere. This makes the daytime temperatures to be much cooler and night-time temperatures to be slightly warmer during the dusty season than the clear sky scenario. Wolkenberg et al., (2020) [167] have correlated the dust opacity with surface temperatures during global dust storms of MY 25, MY 28, and MY 34. He has shown that during daytime, the surface temperature is anti-correlated with dust opacity, whereas, during night-time it is directly correlated. Wolkenberg et al., (2020) [167] have also shown that atmospheric temperature also increases as the

dust opacity increases. Smith, (2004) [16] in figure 6 shows that in presence of a global dust storm the temperature of the whole atmosphere is high as compared to the nonglobal dust storm year. This also proves that the presence of dust in the atmosphere can alter the temperature profile of the atmosphere. These temperature variations can affect the global circulations too. Dust can also affect the atmosphere, by acting as a loss agent for different chemical species such as water cluster ions [66]. Hence understanding the processes by which dust can enter or leave the atmosphere becomes essential. As already stated in previous sections, the dust is lifted into the atmosphere from the ground by near surface winds or convective vortices. These dust are then transported throughout the atmosphere by atmospheric dynamics. Near surface winds cannot lift small dust particles, whereas the dust devils can lift these particles. The small dust particles can remain in suspension for longer times, thus affecting the atmospheric phenomena.

The objective of my thesis is to study one of the dust lifting process, the "dust devil", and provide quantitative investigations of its characteristics in the Martian atmosphere. In chapter 2 [168], we discuss few analytical solutions which are generally used to model for a vortex system in the atmosphere. We then use the Navier Stokes (NS) equation and the continuity equation to determine the mean (with respect to time) tangential wind velocity in a cylindrical co-ordinate system within the surface layer of a planetary atmosphere. For this, we utilize Martian surface layer properties for the theoretical derivation of our solution. In chapter 3 (manuscript under review), we study the characteristics of Martian dust devils using observations from Curiosity rover on Mars. We identify convective vortices by looking at the sudden depression in the

pressure data and also check how many of them show a simultaneous drop in UV flux. Using this, we can determine the possibility that a convective vortex can become a dust devil. In chapter 4 (manuscript under review), we discuss the numerical simulation of dust distribution within a steady-state dust devil. We numerically solve the equations of motion for dust particles to determine their velocity inside the dust devil. Then we use this particle's velocity to determine the spatial distribution of particles inside the dust devil using the continuity equation. In chapter 5, we discuss the electric field generated inside a dust devil and its evolution with time. We develop a two-dimensional cylindrically symmetric model to determine the electric field magnitude inside a dust devil using Poisson's equation, Gauss's law, and charge relaxation equation. This is important in context of lightning generation on Mars. In chapter 6, we summarize the results and conclude the thesis with a discussion about the relevant future research direction.

Chapter 2 Analytical Solutions to Convective Vortices

The planetary boundary layer (PBL) mediates interactions between the surface and the free atmosphere, as discussed in chapter 1. In the Martian PBL, the surface forcings can create convective vortices, which if strong enough to lift dust are called dust devils. A brief description on the generation of convective vortices is given in chapter 1. The winds guiding the vortices are rotating in nature and a knowledge about it is essential to understand the vortex dynamics. The vortex winds are a solution to the Navier-Stokes (NS) equation, which can be considered as Newton's second law of motion for viscous flow in fluids and represents momentum conservation. However, NS equation becomes very complicated to solve even for very simple configurations, necessitating complex numerical solutions for flows like that of a vortex system [169-171]. To avoid such complex computer simulations, meaningful physics of vortices can be understood by using different boundary conditions to find analytical vortex solutions to NS equation. In this chapter, we will discuss these analytical solutions which are generally used as models for vortex systems in the atmosphere. We will then describe another analytical solution that we have developed for the NS equation which helps us to determine the velocity of mean (with respect to time) tangential wind velocity within the surface layer of a planetary atmosphere. Though we have obtained this solution by

utilizing Martian surface layer properties, our results remain valid for any planetary surface layer as long as all of our assumptions are valid.

2.1 Governing equations in Planetary Boundary Layer flows

As already described in chapter 1, the Planetary Boundary Layer (PBL) is directly influenced by atmosphere-surface exchanges of momentum, energy, and mass, in the timescales of an hour or less, making the PBL highly turbulent [38–40]. Thus, in PBL there is a fast temporal evolution of the wind velocity, atmospheric pressure, density and temperature of the surroundings. To describe the incompressible flow within an atmosphere we use three basic equations of fluid dynamics: equation of state, conservation equation for mass, and conservation equation for momentum. A brief description of these equations are given in this section.

2.1.1 Equation of state (Ideal gas law)

An equation of state is defined as a thermodynamic equation relating state variables describing the state of matter under a given set of physical conditions, such as pressure, volume, and temperature. In the atmosphere, the densities of gases are related to their temperature and pressure using ideal gas approximation. Thus, the ideal gas law describing the state of gases in the boundary layer is given as,

$$p = \rho R_a T \qquad \dots (2.1)$$

where, p is the pressure, ρ is the density of the air, R_a is the gas constant per unit mass of air, and T is the absolute temperature. The ideal gas law calculates the behavior of any gas under standard conditions of temperature and pressure.

2.1.2 Conservation of mass (Continuity equation)

A continuity equation describes the transport of air mass and its conservation during the process. The continuity equation simply states that the rate of flow of any gas (mass) inside a given volume is equal to the rate of flow of gas (mass) outside of the same volume plus its accumulation (or loss) in the volume. The differential form of continuity equation, in absence of any source term in the given volume, is given as,

$$\vec{\nabla} \cdot \left(\rho \vec{u}\right) + \frac{\partial \rho}{\partial t} = 0 \qquad \dots (2.2)$$

where, ρ is the density of the air, and \vec{u} is the vector field of gas flow velocity. The divergence part of the equation (2.2) represents the flux of the gas flow, i.e., the difference in rate of flow in and out of the system. The time derivative part of the equation (2.2) represents the accumulation or loss of mass in the volume (negative sign would represent loss). The above equation can be written in another form involving total (material) derivative in which we consider the fluid motion in terms of moving blobs. The continuity equation in terms of the total derivative is,

$$\rho\left(\vec{\nabla}.\vec{u}\right) + \frac{D\rho}{Dt} = 0 \qquad \dots (2.3)$$

where, the total derivative is $\frac{D}{Dt} \equiv \frac{\partial}{\partial t} + \vec{u} \cdot \vec{\nabla}$ and is used to convert between equations (2.2) and (2.3) [172]. The PBL flows are assumed to be incompressible by considering that the density remains constant, i.e., independent of space and time. Thus, the continuity equation reduces to,

$$\vec{\nabla}. \vec{u} = 0 \qquad \dots (2.4)$$

Equation (2.4) means that the flux of gas flow is conserved. Hence, a change in pressure in the atmosphere will be accounted by the change in the gas velocity and not by its density variation.

2.1.3 Conservation of momentum (Navier-Stokes equation)

The circulation of wind in the atmosphere is driven by the pressure gradient developed between two locations, incoming energy from the sun, and the rotation of the planet. The Navier-Stokes (NS) equation describe the motion of viscous fluid and is a formulation of the conservation of momentum of the system, hence called momentum equation also. The NS equation are obtained from Newton's second law, where we consider that a combination of pressure gradient force, viscous force, gravitational force, and Coriolis force are acting together on a moving blob of gas. The NS equation for incompressible flow is written as,

$$\frac{D\vec{u}}{Dt} = -\frac{1}{\rho}\nabla p - 2\left(\vec{\Omega}\times\vec{u}\right) - g\,\hat{k} + \nu\,\nabla^2\,\vec{u} \qquad \dots (2.5)$$

where, ρ is the density of the air, ∇p is the pressure gradient in the atmosphere, $\overline{\Omega}$ is the angular rotation velocity of any planet, \vec{u} is the vector field of gas flow velocity, $g\hat{k}$ is the gravitational force in vertical (z) direction, and ν is the kinematic viscosity of the air (= dynamic viscosity (μ)/density of air (ρ)). Though the NS equation represent winds in any part of an atmosphere, modeling the wind profiles in the boundary layer is challenging [40, 173]. A two dimensional set of NS equation reduces from an elliptical equation to a parabolic equation in the boundary layer under certain assumptions, which is easier to solve [174, 175].

2.2 Analytical solutions to NS equation

A significant amount of dust loading occurs during dust storms in the Martian atmosphere. However, the mechanism of dust lifting is still debatable. One mechanism could be, for example, vortical eruptions resulting from instabilities induced by a convective vortex, which could lead to concentrated vortices or dust devils [176, 177]. Events like dust-devils occur very frequently in the Martian climate system [124, 126, 130, 133, 142]. Therefore, accurate estimation of wind velocity in such a climate system is very important for understanding the system's dynamical behavior. There are many analytical solutions in the literature which are used to study the atmospheric vortex phenomenon like hurricanes, tornadoes, etc. These theoretical vortices with an analytical solution are mainly Rankine vortex, Q vortex, Burgers vortex, Sullivan's vortex and Lamb Oseen vortex. The Rankine, Q vortex and Lamb Oseen vortex solutions are an approximate solution to the NS equation using few boundary layer approximations [178]. The Rankine vortex is a simple analytical model of a vortex in a viscous fluid whereas, Q vortex has its solution dependent on empirically determined constants. The Burgers and Sullivan's vortex solutions are the exact solution to the NS equation [178, 179]. The observations related to Martian atmospheric vortices are best matched by Burgers vortex and Rankine vortex [137]. The tangential velocity component of Q vortex and Lamb Oseen vortex are identical to the Burgers vortex solution of tangential velocity component. Whereas the Sullivan's vortex solution is time dependent but is identical to Burgers solution and even coincides with it as $r \rightarrow \infty$ [179]. Therefore, in this section, we will discuss about the dynamics of vortices, which are presented by the Rankine Vortex and Burgers vortex. These kinds of solution can provide a thorough physical understanding of this phenomena, serve as the testing bed of the accuracy of approximate approaches, and as the basic flow in the stability analyses [178].

2.2.1 The Rankine vortex

The Rankine vortex is a simple model describing only the tangential velocity of the vortex, hence they are also called a stretch-free columnar vortex. It is a fluid flow having radial symmetry and is valid for inviscid flow. It corresponds to a piecewise continuous solution of the NS equation [180]. Its definition is natural in a cylindrical co-ordinate system (r, θ ,z), where the symmetry axis is the z-axis with the r-axis and the θ -axis lying on the plane normal to the z-axis. The Rankine vortex is often called Rankine combined vortex for the reason that it has two separate flow fields. The interior flow field (core) involves only the tangential velocity, which increases linearly with radius from zero along the central axis to a maximum value at a radius (R). Thus this

region rotates like a solid body even though it is fluid. The outer flow (tail) is also purely tangential with the maximum velocity at radius R. The velocity declines inversely with radius from this point outward. The mathematical description of the Rankine combined vortex is,

$$u_{r} = 0$$

$$u_{\theta} = \begin{cases} V_{R} \frac{r}{R}, & \text{if } r \leq R \\ V_{R} \frac{R}{r}, & \text{if } r > R \end{cases}$$
$$u_{z} = 0 \qquad \dots (2.6)$$

where, u_r , u_{θ} and u_z are the components of vortex velocities, V_R is the maximum flow intensity, r is the radial co-ordinate, and R is the radius of the vortex core. Figure 2.1 represents a schematic diagram of the tangential velocity for Rankine vortex. The tangential velocity linearly increases as we move towards the boundary of the vortex and then decreases rapidly as 1/r form.



Figure 2.1: Schematic diagram representing the variation of tangential velocity for the Rankine vortex.

One of the main features of the Rankine vortex is its vorticity field. The vorticity, which is given as $\vec{\omega} = \vec{\nabla} \times \vec{u}$, is constant in the inner part of the vortex. It is a function of the maximum flow velocity (V_R) and the vortex core size (R) only, which is given as $\vec{\omega} = 2 \frac{V_R}{R} \hat{z}$ for $r \leq R$. In the outer region of the vortex, the flow has no vorticity at all. The pressure difference between inside and outside of the Rankine vortex, i.e., the pressure drop within a vortex of Rankine type, is [179, 181, 182],

$$\Delta p = \rho V_R^2 \qquad \dots (2.7)$$

where, ρ is the density of air in the atmosphere.

2.2.2 The Burgers vortex

The Burgers vortex is an exact solution to the NS equation governing viscous flow. Similar to the Rankine vortex, the flow for the Burgers vortex is also described in a cylindrical co-ordinate system (r, θ ,z) having azimuthal symmetry. The mathematical description of the Burgers vortex assuming axial symmetry is,

$$u_{r} = -\frac{1}{2}\alpha r$$

$$u_{\theta} = \frac{\Gamma}{2\pi r} \left[1 - \exp\left(-\frac{\alpha r^{2}}{4v}\right) \right]$$

$$u_{z} = \alpha z \qquad \dots (2.8)$$

where, u_r , u_{θ} and u_z are the components of vortex velocities, $\alpha(>0)$ is the strength of suction, Γ is the circulation strength of the vortex, ν is the viscosity of the atmosphere, r is the radial co-ordinate and z is the axial co-ordinate. Since the axial velocity is dependent on z, vortex stretching occurs in the Burgers vortex. The radial velocity tends to concentrate vorticity around the symmetry axis (z-axis), whereas the viscous diffusion tends to spread the vorticity. This balance leads to the generation of a steady solution to the vortex flow. Figure 2.2 shows a schematic diagram representing vortex stretching along its axial axis for the Burgers vortex [183].



Figure 2.2: Schematic diagram representing vortex stretching along its axial axis for the Burgers vortex [Image credit: Jamil and Shah, 2017 [183]].

The vorticity in the Burgers vortex is given as $\vec{\omega} = \frac{\alpha \Gamma}{4\pi \nu} exp\left(-\frac{\alpha r^2}{4\nu}\right)\hat{z}$. This shows that the vorticity of the Burgers vortex reduces exponentially as we move away from the center of the vortex, proving that vortex strength diminishes outside the core. The pressure drop in the core of the Burgers vortex is [51, 184],

$$\Delta p = -\frac{\rho \left(\Gamma / 2\pi\right)^2}{2r^2}, \qquad \text{for } r \ge r_* \qquad \dots (2.9)$$

where, r_* is the radius of the vortex at which tangential velocity is the maximum, and ρ is the density of air in the atmosphere. Burgers vortex model, which includes stretching of the vortex, is widely used to model turbulent eddies in the atmosphere.

Both Rankine and Burgers vortex analytical solutions are compared with the observations, laboratory experiments and numerical simulation of vortices, which shows reasonable agreement among them [48, 50, 185–192]. The trend of tangential velocity for both Rankine vortex and Burgers vortex are the same. However, unlike Rankine vortex, the radial and vertical velocity components of Burgers vortex are non-zero. Also, unlike Rankine vortex, an inward radial flow in Burgers vortex leads to vertical vorticity enhancement in a narrow column around the symmetry axis [180, 193]. Figure 2.3 shows the variation of tangential velocity for both Rankine vortex and Burgers vortex by considering all parameters to be unity. We observe that the transition of velocity at the vortex boundary is smooth for Burgers' solution in comparison to the Rankine's solution, which shows a discontinuity at the vortex core (r = R). This makes the Burgers vortex model more suitable and realistic in interpreting the vortices in the atmosphere [180].



Figure 2.3: The variation of tangential velocity for both Rankine vortex and Burgers vortex.

2.3 Tangential winds for a vortex derived from NS equation

Most often, a cylindrical system is used for experiments or simulations of vortex systems [51, 194, 195]. Therefore, we consider a cylindrical vortex system to estimate the wind velocity variation with radial distance from the vortex center, and altitude. A cylindrical co-ordinate system is a three-dimensional co-ordinate system that specifies point positions by the distance from a chosen reference axis, the direction from the axis relative to a chosen reference direction, and the distance from a chosen reference plane perpendicular to the axis [196]. The expression for the mean horizontal velocity of wind in the Cartesian co-ordinate system provides a good approximation for atmospheric wind movement in the logarithmic layer of any planet. However, this expression fails

to accurately determine the tangential velocity of wind for a vortex system such as dust devil, tornado, and storms. The objective of this section is to derive the equation of mean tangential velocity of wind in a cylindrical co-ordinate system (for the planetary surface layer) using NS and continuity equations. Our co-ordinate system coincides with the center of the vortex system at the surface. We assume our vortex system to be in steady-state. In the steady-state condition, the structure of a vortex system does not undergo any change with time. The velocity field around the vortex is always normal to both the symmetry axis: the vertical axis 'z' and the radial vector 'r'. We do not consider any effect of the translational motion of the vortex in the estimation of our velocity. We will also compare our results with observed tangential velocities of the vortex on Earth for validation of our results in the current section.

2.3.1 Mathematical formulation of the problem

Figure 2.4 shows a schematic of a cylindrical vortex system with mean tangential wind flowing in an anti-clockwise direction.



Figure 2.4: Cylindrical vortex system with mean tangential wind (\overline{u}_{θ}) flowing in anticlockwise direction.
The NS equation in rotational frame of reference (tensorial notation) for incompressible fluid flow, is given as [197, 198]:

$$\frac{\partial u_i}{\partial t} + v_j \frac{\partial u_i}{\partial x_j} = -\frac{1}{\rho} \frac{\partial p}{\partial x_i} + v \frac{\partial^2 u_i}{\partial x_j^2} - g \,\delta_{i3} - 2\Omega \varepsilon_{ijk} \eta_j u_k \qquad \dots (2.10)$$

where, u_i is the component of fluid velocity, ρ is the density of the fluid, ν is the kinematic viscosity of the fluid (= dynamic viscosity (μ) / density of fluid (ρ)), p is the pressure of the surrounding, g is the acceleration due to the gravity of the planet and Ω is the angular rotation speed of the planet. The continuity equation for an incompressible flow of fluid (tensorial notation) is given as:

$$\frac{\partial u_i}{\partial x_i} = 0 \qquad \dots (2.11)$$

In the cylindrical co-ordinate system, the equations of atmospheric flow are expressed with components (u_r, u_θ, u_z) of velocity vector \vec{u} . Therefore, the continuity equation can be expressed as:

$$\frac{\partial u_r}{\partial r} + \frac{u_r}{r} + \frac{1}{r} \frac{\partial u_{\theta}}{\partial \theta} + \frac{\partial u_z}{\partial z} = 0 \qquad \dots (2.12)$$

Thus, the NS equation (equation (2.5)) in a cylindrical co-ordinate system is given as:

• The *r*-component:

$$\frac{\partial u_r}{\partial t} + u_r \frac{\partial u_r}{\partial r} + \frac{u_{\theta}}{r} \frac{\partial u_r}{\partial \theta} + u_z \frac{\partial u_r}{\partial z}$$

$$= -\frac{1}{\rho} \left(\frac{\partial P}{\partial r} \right) + \frac{u_{\theta}^2}{r} + v \left(\frac{\partial^2 u_r}{\partial r^2} + \frac{\partial^2 u_r}{\partial z^2} + \frac{1}{r^2} \frac{\partial^2 u_r}{\partial \theta^2} - \frac{2}{r^2} \frac{\partial u_{\theta}}{\partial \theta} + \frac{1}{r} \frac{\partial u_r}{\partial r} - \frac{u_r}{r^2} \right)$$
...(2.13)

• The θ -component:

$$\frac{\partial u_{\theta}}{\partial t} + u_{r} \frac{\partial u_{\theta}}{\partial r} + \frac{u_{\theta}}{r} \frac{\partial u_{\theta}}{\partial \theta} + u_{z} \frac{\partial u_{\theta}}{\partial z}$$

$$= -\frac{1}{\rho} \frac{1}{r} \left(\frac{\partial P}{\partial \theta} \right) - \frac{u_{r} u_{\theta}}{r} + v \left(\frac{\partial^{2} u_{\theta}}{\partial r^{2}} + \frac{\partial^{2} u_{\theta}}{\partial z^{2}} + \frac{1}{r^{2}} \frac{\partial^{2} u_{\theta}}{\partial \theta^{2}} + \frac{2}{r^{2}} \frac{\partial u_{r}}{\partial \theta} + \frac{1}{r} \frac{\partial u_{\theta}}{\partial r} - \frac{u_{\theta}}{r^{2}} \right) \qquad \dots (2.14)$$

• The *z*-component:

$$\frac{\partial u_z}{\partial t} + u_r \frac{\partial u_z}{\partial r} + \frac{u_\theta}{r} \frac{\partial u_z}{\partial \theta} + u_z \frac{\partial u_z}{\partial z}$$

$$= -\frac{1}{\rho} \left(\frac{\partial P}{\partial z} \right) + g \hat{z} + v \left(\frac{\partial^2 u_z}{\partial r^2} + \frac{\partial^2 u_z}{\partial z^2} + \frac{1}{r^2} \frac{\partial^2 u_z}{\partial \theta^2} + \frac{1}{r} \frac{\partial u_z}{\partial r} \right) \qquad \dots (2.15)$$

According to the atmospheric boundary layer theory, the Coriolis force near the surface layer can be neglected [42]. Again, if the horizontal scale of a disturbance is small enough, the Coriolis force may be neglected compared to the pressure gradient force and the centrifugal force. The Rossby number (*Ro*), Ro = V/f L, where *V* is the wind-speed, $f = 2\Omega sin\phi$ is the Coriolis parameter (Ω is the angular frequency of planetary rotation and ϕ is the latitude), and *L* is the length scale of vortex, determines the relative significance of various forces with each other. For Mars surface layer winds, $V = 10 ms^{-1}$, L = 100 m, and $f = 10^{-4} s^{-1}$ leads to $Ro \approx 10^3$, which implies that the Coriolis force can be neglected as compared to other forces. Moreover, atmospheric motions are usually modeled within the shallow-fluid approximation. This simplifies the three-

dimensional spherical geometry, and for dynamical consistency, the Coriolis force is neglected. We consider the incompressible flow of the fluid, which is an important condition for applying Boussinesq approximation. The Boussinesq approximation ignores density differences except where they appear in terms multiplied by gravity 'g'. The idea behind the Boussinesq approximation is to restrict the analysis to that of a system whose overall background density and temperature do not vary much around their mean values. The mathematical form for pressure, density, and temperature using Boussinesq approximation [38, 199] can be given as:

$$P = P_o(z) + p(\vec{r}), \quad \rho = \rho_o(z) + \rho'(\vec{r}), \quad T = T_o(z) + T'(\vec{r}) \qquad \dots (2.16)$$

where, $P_0(z)$, $\rho_0(z)$, and $T_0(z)$ are the time-averaged values of pressure, density and temperature of the fluid respectively, and $p(\vec{r})$, $\rho'(\vec{r})$, $T'(\vec{r})$ are the fluctuations in this time-averaged values of pressure, density, and temperature of the fluid, respectively.

The adiabatic lapse rate for the atmosphere is defined as the rate of change of temperature of an air parcel with altitude:

$$\frac{\partial T_0}{\partial z} = -\frac{g}{C_P} \qquad \dots (2.17)$$

The hydrostatic equilibrium equation is:

$$\frac{\partial P_0}{\partial z} = -g\,\rho_0 \qquad \dots (2.18)$$

Using equations (2.16) to (2.18), the three components of the NS equation can be written as:

(a) The *r*-component:

$$\frac{\partial u_r}{\partial t} + u_r \frac{\partial u_r}{\partial r} + \frac{u_{\theta}}{r} \frac{\partial u_r}{\partial \theta} + u_z \frac{\partial u_r}{\partial z}$$

$$= -\frac{1}{\rho_o} \left(\frac{\partial p}{\partial r}\right) + \frac{u_{\theta}^2}{r} + v \left(\frac{\partial^2 u_r}{\partial r^2} + \frac{\partial^2 u_r}{\partial z^2} + \frac{1}{r^2} \frac{\partial^2 u_r}{\partial \theta^2} - \frac{2}{r^2} \frac{\partial u_{\theta}}{\partial \theta} + \frac{1}{r} \frac{\partial u_r}{\partial r} - \frac{u_r}{r^2}\right)$$
...(2.19)

(b) The θ -component:

$$\frac{\partial u_{\theta}}{\partial t} + u_{r} \frac{\partial u_{\theta}}{\partial r} + \frac{u_{\theta}}{r} \frac{\partial u_{\theta}}{\partial \theta} + u_{z} \frac{\partial u_{\theta}}{\partial z}$$

$$= -\frac{1}{\rho_{o}} \frac{1}{r} \left(\frac{\partial p}{\partial \theta}\right) - \frac{u_{r} u_{\theta}}{r} + v \left(\frac{\partial^{2} u_{\theta}}{\partial r^{2}} + \frac{\partial^{2} u_{\theta}}{\partial z^{2}} + \frac{1}{r^{2}} \frac{\partial^{2} u_{\theta}}{\partial \theta^{2}} + \frac{2}{r^{2}} \frac{\partial u_{r}}{\partial \theta} + \frac{1}{r} \frac{\partial u_{\theta}}{\partial r} - \frac{u_{\theta}}{r^{2}}\right) \dots (2.20)$$

(c) The *z*-component:

$$\frac{\partial u_{z}}{\partial t} + u_{r} \frac{\partial u_{z}}{\partial r} + \frac{u_{\theta}}{r} \frac{\partial u_{z}}{\partial \theta} + u_{z} \frac{\partial u_{z}}{\partial z}$$

$$= -\frac{1}{\rho_{o}} \left(\frac{\partial p}{\partial z}\right) - g \frac{\rho'}{\rho_{o}} + v \left(\frac{\partial^{2} u_{z}}{\partial r^{2}} + \frac{\partial^{2} u_{z}}{\partial z^{2}} + \frac{1}{r^{2}} \frac{\partial^{2} u_{z}}{\partial \theta^{2}} + \frac{1}{r} \frac{\partial u_{z}}{\partial r}\right) \qquad \dots (2.21)$$

Applying linearity, i.e., $\frac{\rho'}{\rho_0} = \frac{-T'}{T_0}$ to equation (2.21), and with further simplification we

get:

$$\frac{\partial u_z}{\partial t} + u_r \frac{\partial u_z}{\partial r} + \frac{u_\theta}{r} \frac{\partial u_z}{\partial \theta} + u_z \frac{\partial u_z}{\partial z}$$

= $-\frac{1}{\rho_0} \left(\frac{\partial p}{\partial z}\right) + g \frac{T}{T_0} + v \left(\frac{\partial^2 u_z}{\partial r^2} + \frac{\partial^2 u_z}{\partial z^2} + \frac{1}{r^2} \frac{\partial^2 u_z}{\partial \theta^2} + \frac{1}{r} \frac{\partial u_z}{\partial r}\right)$...(2.22)

Chain rule for some partial differential terms in equations (2.19), (2.20) and (2.22) is as follows:

$$u_{r}\frac{\partial u_{r}}{\partial r} + \frac{u_{\theta}}{r}\frac{\partial u_{r}}{\partial \theta} + u_{z}\frac{\partial u_{r}}{\partial z} = \frac{\partial (u_{r}u_{z})}{\partial z} + \frac{1}{r}\frac{\partial (r u_{r}u_{r})}{\partial r} + \frac{1}{r}\frac{\partial (u_{r}u_{\theta})}{\partial \theta} \qquad \dots (2.23)$$

$$u_{r}\frac{\partial u_{\theta}}{\partial r} + \frac{u_{\theta}}{r}\frac{\partial u_{\theta}}{\partial \theta} + u_{z}\frac{\partial u_{\theta}}{\partial z} = \frac{\partial (u_{\theta}u_{z})}{\partial z} + \frac{1}{r}\frac{\partial (r u_{r}u_{\theta})}{\partial r} + \frac{1}{r}\frac{\partial (u_{\theta}u_{\theta})}{\partial \theta} \qquad \dots (2.24)$$

$$u_{r}\frac{\partial u_{z}}{\partial r} + \frac{u_{\theta}}{r}\frac{\partial u_{z}}{\partial \theta} + u_{z}\frac{\partial u_{z}}{\partial z} = \frac{\partial (u_{z}u_{z})}{\partial z} + \frac{1}{r}\frac{\partial (ru_{r}u_{z})}{\partial r} + \frac{1}{r}\frac{\partial (u_{z}u_{\theta})}{\partial \theta} \qquad \dots (2.25)$$

After applying chain rule (equation (2.23 - 2.25)), the three components of the NS equation (equation (2.19 - 2.22)) will simplify to:

(a) The *r*-component:

$$\frac{\partial u_r}{\partial t} + \frac{1}{r} \frac{\partial (r u_r u_r)}{\partial r} + \frac{1}{r} \frac{\partial (u_r u_\theta)}{\partial \theta} + \frac{\partial (u_r u_z)}{\partial z}$$
$$= -\frac{1}{\rho_o} \left(\frac{\partial p}{\partial r}\right) + \frac{u_\theta^2}{r} + v \left(\frac{\partial^2 u_r}{\partial r^2} + \frac{\partial^2 u_r}{\partial z^2} + \frac{1}{r^2} \frac{\partial^2 u_r}{\partial \theta^2} - \frac{2}{r^2} \frac{\partial u_\theta}{\partial \theta} + \frac{1}{r} \frac{\partial u_r}{\partial r} - \frac{u_r}{r^2}\right) \dots (2.26)$$

(b) The θ -component:

$$\frac{\partial u_{\theta}}{\partial t} + \frac{1}{r} \frac{\partial (r u_{r} u_{\theta})}{\partial r} + \frac{1}{r} \frac{\partial (u_{\theta} u_{\theta})}{\partial \theta} + \frac{\partial (u_{\theta} u_{z})}{\partial z}$$

$$= -\frac{1}{\rho_{o}} \frac{1}{r} \left(\frac{\partial p}{\partial \theta}\right) - \frac{u_{r} u_{\theta}}{r} + \nu \left(\frac{\partial^{2} u_{\theta}}{\partial r^{2}} + \frac{\partial^{2} u_{\theta}}{\partial z^{2}} + \frac{1}{r^{2}} \frac{\partial^{2} u_{\theta}}{\partial \theta^{2}} + \frac{2}{r^{2}} \frac{\partial u_{r}}{\partial \theta} + \frac{1}{r} \frac{\partial u_{\theta}}{\partial r} - \frac{u_{\theta}}{r^{2}}\right) \dots (2.27)$$

(c) The *z*-component:

$$\frac{\partial u_z}{\partial t} + \frac{1}{r} \frac{\partial \left(r \, u_r u_z\right)}{\partial r} + \frac{1}{r} \frac{\partial \left(u_z u_\theta\right)}{\partial \theta} + \frac{\partial \left(u_z u_z\right)}{\partial z}$$
$$= -\frac{1}{\rho_0} \left(\frac{\partial p}{\partial z}\right) + g \frac{T}{T_0} + v \left(\frac{\partial^2 u_z}{\partial r^2} + \frac{\partial^2 u_z}{\partial z^2} + \frac{1}{r^2} \frac{\partial^2 u_z}{\partial \theta^2} + \frac{1}{r} \frac{\partial u_z}{\partial r}\right) \qquad \dots (2.28)$$

According to the Reynolds-averaged treatment [200, 201], an instantaneous quantity is decomposed into its time-averaged and fluctuating quantities, i.e., $\vec{u}(\vec{r}, t) =$

 $\vec{u}(\vec{r},t) + \vec{u}'(\vec{r},t)$. This treatment is primarily used to describe turbulent flows. Using Reynolds treatment, the continuity equation (equation (2.12)) can be re-written as:

$$\frac{1}{r}\frac{\partial(r\bar{u}_r)}{\partial r} + \frac{1}{r}\frac{\partial\bar{u}_{\theta}}{\partial\theta} + \frac{\partial\bar{u}_z}{\partial z} = 0 \qquad \dots (2.29)$$

An approximation to the Reynolds number can be given as $R_e \approx VL/v$ [45], where V and L are typical velocity scale and typical length scale for a flow, respectively and vis the kinematic viscosity of the medium (air). The Reynolds number is the ratio of inertial forces to viscous forces and measures how turbulent the flow is. Low Reynolds number flows are laminar, while higher Reynolds number flows are turbulent. For Mars surface layer winds, $V \approx 10 \text{ ms}^{-1}$, $L \approx 100 \text{ m}$, and $v \approx 10^{-3} \text{ m}^2 \text{s}^{-1}$ [40], it follows that R_e $\approx 10^6$. A large Reynolds number indicates that the wind flows in Martian surface layers are almost always turbulent. Therefore, neglecting the viscous terms in equations (2.26) to (2.28) (since we are dealing with a turbulent region of the atmosphere), the three components of Reynolds-averaged NS equation will be reduced to:

$$\frac{\partial \overline{u}_{r}}{\partial t} + \overline{u}_{r} \frac{\partial \overline{u}_{r}}{\partial r} + \frac{\overline{u}_{\theta}}{r} \frac{\partial \overline{u}_{r}}{\partial \theta} + \overline{u}_{z} \frac{\partial \overline{u}_{r}}{\partial z}
= -\frac{1}{\rho_{o}} \frac{\partial \overline{p}}{\partial r} - \frac{\partial \left(\overline{u'_{r}u'_{z}}\right)}{\partial z} - \frac{\partial \left(\overline{u'_{r}u'_{r}}\right)}{\partial r} - \frac{1}{r} \frac{\partial \left(\overline{u'_{r}u'_{\theta}}\right)}{\partial \theta} - \frac{1}{r} \frac{\overline{u'_{r}u'_{r}}}{u'_{r}u'_{r}} + \frac{1}{r} \frac{\overline{u_{\theta}u_{\theta}}}{u_{\theta}} + \frac{1}{r} \overline{u'_{\theta}u'_{\theta}} \qquad \dots (2.30)$$

$$\frac{\partial \overline{u}_{\theta}}{\partial t} + \overline{u}_{r} \frac{\partial \overline{u}_{\theta}}{\partial r} + \frac{\overline{u}_{\theta}}{r} \frac{\partial \overline{u}_{\theta}}{\partial \theta} + \overline{u}_{z} \frac{\partial \overline{u}_{\theta}}{\partial z}
= -\frac{1}{\rho_{o}} \frac{\partial \overline{p}}{\partial \theta} - \frac{\partial \left(\overline{u_{\theta}' u_{z}'}\right)}{\partial z} - \frac{\partial \left(\overline{u_{\theta}' u_{z}'}\right)}{\partial r} - \frac{1}{r} \frac{\partial \left(\overline{u_{\theta}' u_{\theta}'}\right)}{\partial \theta} - \frac{1}{r} \overline{u_{r}' u_{\theta}'} - \frac{1}{r} \overline{u_{r}' u_{\theta}'} - \frac{1}{r} \overline{u_{r}' u_{\theta}'} \qquad \dots (2.31)$$

$$\frac{\partial \overline{u}_{z}}{\partial t} + \overline{u}_{r} \frac{\partial \overline{u}_{z}}{\partial r} + \frac{\overline{u}_{\theta}}{r} \frac{\partial \overline{u}_{z}}{\partial \theta} + \overline{u}_{z} \frac{\partial \overline{u}_{z}}{\partial z}
= -\frac{1}{\rho_{o}} \frac{\partial \overline{p}}{\partial z} - \frac{\partial \left(\overline{u'_{z}u'_{z}}\right)}{\partial z} - \frac{\partial \left(\overline{u'_{r}u'_{z}}\right)}{\partial r} - \frac{1}{r} \frac{\partial \left(\overline{u'_{z}u'_{\theta}}\right)}{\partial \theta} - \frac{1}{r} \frac{u'_{r}u'_{z}}{u'_{r}u'_{z}} \qquad \dots (2.32)$$

The azimuthal velocity (tangential velocity) determines the rotational velocity of a rotating vortex and the amount of dust it can lift. Since we have considered the velocity field to be normal to both 'z' and 'r' axis, the velocity field is parallel to the $\hat{\theta}$ direction. Therefore the net velocity field would be, $\vec{u} = u_{\theta}\hat{e_{\theta}}$. Now, we are interested in determining the variation of azimuthal velocity with the height and size of the vortex. Therefore, we consider that θ -component is in the direction of mean horizontal velocity, z is in the direction of mean vertical velocity, and the ground is homogeneous with even roughness. Since Coriolis force has been already neglected, that will lead to the cyclostrophic balance in an atmospheric vortex. The cyclostrophic balance occurs within a vortex when the horizontal pressure gradient, which is acting inwards, is counter-balanced by the centrifugal acceleration of the winds in an outward direction. Thus, cyclostrophic balance [199, 202] will satisfy:

$$\frac{V^2}{r} \approx -\frac{1}{\rho} \frac{\partial p}{\partial n} \qquad \dots (2.33)$$

where, n is normal to the direction of flow. After applying cyclostrophic balance, equation (2.31) can be simplified to:

$$\frac{\partial \overline{u}_{\theta}}{\partial t} = -\frac{1}{\rho_o} \frac{\partial \overline{p}}{\partial \theta} - \frac{\partial \left(\overline{u_{\theta}' u_z'}\right)}{\partial z} - \frac{2}{r} \overline{u_r' u_{\theta}'} \qquad \dots (2.34)$$

Since we have taken a cylindrical symmetry, we do not expect any change in pressure in the azimuthal direction. Hence, equation (2.34) simplifies to,

$$\frac{\partial \overline{u}_{\theta}}{\partial t} = \frac{1}{\rho_o} \frac{\partial \left(-\rho_o \overline{u_{\theta}' u_z'}\right)}{\partial z} - \frac{2}{r} \overline{u_r' u_{\theta}'} \qquad \dots (2.35)$$

In a steady-state scenario, equation (2.35) can be written as:

$$\frac{1}{\rho_o} \frac{\partial \left(-\rho_o \overline{u_{\theta}' u_z'}\right)}{\partial z} - \frac{2}{r} \overline{u_r' u_{\theta}'} = \frac{1}{\rho_o} \frac{\partial \tau_{z\theta}}{\partial z} + \frac{2\tau_{r\theta}}{r\rho_o} = 0 \qquad \dots (2.36)$$

where $\tau_{z\theta}$ and $\tau_{r\theta}$ are the components of Reynolds shear stress [38, 45].

Reynolds shear stress $(\tau_{r\theta})$ is described using the expression $-\rho \overline{u_r} u_{\theta}$. Reynolds stress only exists when the fluid is in turbulent motion. The Reynolds shear stress deals with turbulent momentum flux which acts like a stress. The momentum flux is transferred to other layers by the fluctuating winds. Hence, the Reynolds stress becomes directly dependent on the velocity of the wind, and not on the position of the vortex. Since $\tau_{r\theta}$ is only a function of velocities, and not co-ordinates; we integrate equation (2.36) (integration limit, $z_0 \rightarrow z$) and get,

$$\tau_{z\theta} = -\frac{2(z-z_o)}{r}\tau_{r\theta} \qquad \dots (2.37)$$

Turbulence shear stress can be given as [38]:

$$\tau = \sqrt{\tau_{r\theta}^2 + \tau_{z\theta}^2} = \tau_{r\theta} \sqrt{1 + \frac{4(z - z_o)^2}{r^2}} \qquad \dots (2.38)$$

Prandtl postulated that the velocity scale of a fluctuation motion is equal to the velocity gradient times the mixing length scale [203]. According to Prandtl's mixing length theory [42], the shear stress can be given by $\tau = \mu_t \frac{\partial \overline{u_\theta}}{\partial z}$; where $\mu_t = \rho_0 l_m^2 \left| \frac{\partial \overline{u_\theta}}{\partial z} \right|$, and $l_m = kz$ is mixing length. Using Prandtl's theory and equation (2.38), we get:

$$\tau_{r\theta} \sqrt{1 + \frac{4(z - z_o)^2}{r^2}} = \rho_o k^2 z^2 \left| \frac{\partial \overline{u}_{\theta}}{\partial z} \right|^2 \qquad \dots (2.39)$$

Using the expression of threshold friction $u_*(=\sqrt{\tau_{r\theta}/\rho_0})$ velocity near the surface [45], and further simplifying equation (2.39), we get:

$$\frac{\partial \overline{u}_{\theta}}{\partial z} = \frac{u_*}{kz} \left(1 + \frac{4(z - z_o)^2}{r^2} \right)^{\frac{1}{4}}, \qquad z \ge z_o \qquad \dots (2.40)$$

where 'r' is the distance from the center of the vortex, and 'z' is the altitude from the surface. Equation (2.40) is the analytical solution (with certain assumptions) of Navier-Stokes equation for estimating the tangential velocity for a vortex system in a planetary surface layer.

2.3.2 Results and discussions

We use the trapezoidal method for numerical integration (integration limit, $z_0 \rightarrow z$) of equation (2.40) to determine tangential wind velocities in Earth's and Mars' surface layers. We assume values of 2 ms⁻¹, 0.4, and 0.01 m for near-surface threshold friction velocity (u_*), von Karman constant (k), and aerodynamic roughness length (z_o) of the surface respectively for a typical Martian surface layer [40, 46, 204, 205]. For

Earth, we assume values of 1 ms⁻¹ and 0.03 m for near-surface threshold friction velocity and aerodynamic roughness length of the surface, respectively [206, 207]. Typical Mars' and Earth's surface layers extend up to 1000 m and 150 m, respectively. Figure 2.5 shows the variation of mean tangential wind velocity with altitude for various radial distances. We observe a reduction by a factor of about 1.5 in mean velocities (at 150 m altitude for both planets) for one order increase (10 m to 100 m) in the radial distance due to inverse square dependency on radial distance. The tangential velocities would be higher as we move closer to the center of the vortex and decreases sharply with increasing distance. The variation in tangential velocities with variation in radial distance increases with increasing altitude and vice-versa.



Figure 2.5: Comparison of tangential wind velocity profiles of Earth (top) and Mars (Bottom) for various radial distances from vortex center.

Figure 2.6 shows the comparison of velocities between Earth and Mars surface layers at a fixed altitude (7 ft (2.13 m)) with respect to the radial distance from the center (0 indicates the center) of the vortex. The velocities on Mars are relatively higher as compared to velocities on Earth due to lower roughness length and higher threshold friction velocity. As we move far away from the center of a rotating vortex, the velocities in both planets' surface layers reach a saturation value, which is the ambient wind speed. This happens because as 'r' increases $(r \rightarrow \infty)$, the second term in equation (2.40) becomes non-significant as compared to the first term, and eventually leads to logarithmic wind profiles without any effect of the vortex. The uncertainties involved with the measurements of roughness length and friction velocities in Martian atmosphere will not significantly affect our estimated tangential wind velocity. These quantities are determined using chamber experiments for Martian atmospheric conditions and also by using Viking landers data [46]. While determining the wind speed using Viking lander dataset an inaccuracy of $\pm 15\%$ is reported [40]. The value of z_o is expected to vary within the range 0.001 to 0.01 m [205]. The von Karman constant (k) is reported to be of the order 0.400 ± 0.011 [204]. If we fix $u_* = 2$ m/s and k = 0.4, and vary the magnitude of z_0 from 0.001 m to 0.01 m, we obtain the magnitude of tangential wind velocity to be 21.469 m/s and 21.467 m/s at 2.13 m altitude for r = 10m. This shows that the uncertainty does not affect the final result much.



Figure 2.6: Comparison of tangential wind velocities on Mars and Earth due to a vortex with respect to the radial distance from the center (0 being the center). Velocities are estimated at 7 ft (2.13 m) altitude from the surface.

2.3.3 Comparison with observed data

Sinclair (1966) [208] measured three cylindrical components of the wind velocity through the base of a dust devil, at 7, 17, and 31 ft above the surface over a flat desert terrain near Tucson, Arizona. From these observations, it follows that the tangential wind velocity typically fluctuates between 10 and 15 ms⁻¹ [50]. For comparison, we theoretically determine the tangential wind velocity using equation (2.40), with roughness length (z_0) = 0.03 m [207], and $u_* = 1 \text{ ms}^{-1}$ [206] at 7, 17, and 31 ft (2.13, 5.18, and 9.45 m) above the surface. We determine the velocities at 5 m and 10 m from the center of the vortex. In general, our theoretical estimates of the tangential

wind velocity (Table 2.1) are well within the range of observed values. At higher altitude and close to vortex center, our values are slightly higher due to various assumptions (or approximations) we made during the mathematical formulation of our derived equation.

Table 2.1: Theoretical tangential wind velocity $(\overline{u_{\theta}})$ determined using equation (2.40) with $z_0=0.03$ m, and $u_*=1$ ms⁻¹. All velocities are in ms⁻¹.

	z = 7 ft (2.13 m)	z = 17 ft (5.18 m)	z = 31 ft (9.45 m)
r = 5 m	10.88	13.78	16.38
r = 10 m	10.73	13.19	15.17

Cyclostrophic balance is always maintained within a vortex. Therefore, we can predict the pressure drop around a vortex using this balance if we have the knowledge of tangential velocity and vice versa. The cyclostrophic balance around a vortex can also be represented as,

$$\Delta p = \frac{V^2}{RT} p_{avg} \qquad \dots (2.41)$$

where, Δp is the pressure drop around the vortex, V is the tangential velocity of the vortex, R is the specific gas constant of the air, T is the background temperature and p_{avg} is the average pressure of the surrounding. Sinclair (1966) [208] measured the

tangential velocity of the dust devil at three heights varying between 10 - 15 m/s and pressure drop varying between 1 - 3 hPa. Using the value of the reported T = 320 K and assuming the typical summertime values for the surface pressure $p_{avg} = 925$ hPa and R = 287 m²s⁻²K⁻¹, we obtain the calculated pressure drop to vary between 1.2 to 2.7 hPa. For Mars, with typical values of T = 250 K, R = 192 m²s⁻²K⁻¹, and $p_{avg} = 700$ Pa and threshold velocity $V \approx 30$ ms⁻¹ [149], we obtain a pressure drop of $\Delta p \approx 13$ Pa.

2.4 Conclusion

In this chapter, we have discussed about the basic equations governing the fluid flow in the atmosphere and few analytical solutions for the vortex system. We have then demonstrated another analytical solution to the NS equation with the help of the continuity equation (equation (2.40)), to derive a simple form for the mean tangential velocity in a cylindrical co-ordinate system. The derived equation represents the dependency of tangential velocity on distance from the center of the cylinder, and the altitude. The equation (2.40) would be useful to estimate the variation of velocity with radial distance from the vortex center. However, as we move further away from the vortex center, the second term in equation (2.40) becomes non-significant, and velocities start following the standard logarithmic profile. We note that equation (2.40) is only valid in the planetary surface layer region of the atmosphere.

The dependency of tangential wind velocity on altitude indicates the increase in velocity as we move higher up in the vortex system. The tangential velocity of the wind also decreases as we move far away from the vortex center. At 100 m altitude, for an

order of magnitude increase in the radial distance, the mean tangential wind velocity drops by about a factor of 1.5 in magnitude. We theoretically estimated tangential velocities for both Earth and Mars surface layers. The velocities on Earth are relatively lower as compared to velocities on Mars due to higher roughness length and lower threshold friction velocity.

A comparison with observed data substantiates the validity and applicability of equation (2.40) for vortex systems in planetary surface layers. Unlike other analytical solutions for the vortex, our solution is dependent on the measureable atmospheric parameters like threshold friction velocity and roughness length of the atmosphere. We believe that this form of the equation can prove very vital for determining the mean tangential wind velocities in a vortex system for a planet like Mars (where vortex systems like dust devils occur frequently). Although we utilize Martian surface layer properties to derive most of our results, the equation (2.40) remains valid for any planetary surface layer as long as all of our assumptions are valid.

Chapter 3 Observational Study of the Convective Vortices

Convective vortices are believed to be an efficient lifting mechanism for dust from the planetary surface into the Martian atmosphere [209]. Such dust lifting vortices also known as dust devils, can be identified in-situ, by landers or rovers through a sudden decrease in surface pressure or through camera images. They can also be identified from images taken by orbiting spacecrafts. The dust devils maintain the background atmospheric haze of the atmosphere, thus influencing the thermal and dynamical state of the atmosphere [52, 57, 61, 62, 131, 138, 142]. Dust devils are hazardous for robotic missions and also human missions, being planned for Mars. Therefore, a proper understanding of the dust devil is important before planning such missions. It is hard to predict the exact location and time of occurrence of dust devils which makes the in-situ observations of dust devils and convective vortices a challenge [210]. In this chapter, we will discuss the methods (in-situ and from orbit) for detection of convective vortices and dust devils in the Martian atmosphere. We have used one of these methods to study different characteristics of convective vortices in Gale Crater (5.4°S, 137.8°E), which will be discussed in detail.

3.1 Observation of dust devils using images

As already mentioned in chapter 1, several missions have sent spacecrafts (orbiters) to Mars which provides us knowledge about the Martian surface and atmosphere from the orbit using several on-board scientific instruments. Dust devils were first imaged from orbit by Viking Orbiter cameras as small bright columns of dust with long shadows [112]. Since then they have been observed through images obtained from other orbiter missions too, like Mars Orbiter Camera (MOC) onboard the Mars Global Surveyor (MGS) [113, 115, 117–119]; Thermal Emission Imaging System (THEMIS) onboard the Mars Odyssey spacecraft [114, 120]; High Resolution Stereo Camera (HRSC) on-board Mars Express (MEX) [121–126]; Mars Color Imager (MARCI), Context Camera (CTX), and High Resolution Imaging Experiment (HiRISE) on-board Mars Reconnaissance Orbiter (MRO) spacecraft [127, 211]; and Mars Color Camera (MCC) on-board Mars Orbiter Mission (MOM) [128]. Images from These orbiters have been used to determine the diameter and height of the dust devils using the "shadow method". One such image of dust devils observed by MCC is shown in figure 3.1 [128].



Figure 3.1: MCC observation of three Dust Devils, marked by yellow arrows, on 7th November 2016 [Image credit: Singh and Arya, 2019 [128]].

The image by MCC, was taken in the visible range $(0.4 - 0.7 \,\mu\text{m})$, during Ls = 256.83° at 12:08:56 UTC on 7th November 2016 (corresponding to MY 33) with 25.52 m resolution at spacecraft altitude of ~490 km in the southern hemisphere of Mars. The shadow method was used with a sun-sensor geometry to determine the height of the dust devils, which varied from ~0.5 – 1.9 km [128]. Observations from several other orbiters estimate sizes of dust devils in the range of a few meters to tens of meters in diameter, and tens of meters to over a few kilometers in height [125, 212].

The orbital images of dust devils suggest, that they generally form almost anywhere – over latitudes ranging from 80°S to 80°N, and from the bottom of the Hellas Basin to the tops of volcanoes [113–115]. In general, dust devils at a particular location can be inferred from the dust devil tracks (DDT) in the images. Whelley and Greeley, (2006) [213] determined the DDT densities for regions centered on Gusev crater (30° S to 0° N; 170° W to 300° W), Ares Vallis (5° N to 35° N; 15° W to 45° W) and for a pole to pole swath (90° S to 90° N; 100° W to 120° W). They obtained the average DDT density in the southern hemisphere to be ~0.6 DDT/km² and in the northern hemisphere to be ~0.06 DDT/km², thus predicting that dust devils are more in the southern hemisphere than in northern hemisphere [213]. Fisher et al., (2005) [118] suggests that DDT might not be a good proxy for dust devil activity, since they have observed many active dust devils in Amazonis but relatively few DDT; and many DDT but no active dust devils in Casius. Considering all these information, some active dust devil regions are identified on Mars from orbital images which include Amazonis Planitia (~30° N, ~190° E) [113, 117, 118, 214], Casius (~40° N, ~90° E) [118], Argyre Planitia (~50° S, ~340° E) [139], and Syria-Claritas (~15° S, ~251° E), and eastern Meridiani just west of Schiaparelli Crater (~3° S, ~17° E) [115]. Moreover, the dust devils are observed in all seasons in both hemispheres, having peak activity (maximum number of dust devils) during the corresponding summer season [115, 117, 118, 142].

The observations of dust devils from orbit are restricted by the fixed local times dictated by a spacecraft's orbit. Due to this, it is not possible to get information on the diurnal variation of dust devil activity. Therefore, the surface-based instruments (landers and rovers) are used to provide information regarding the diurnal and seasonal variability of dust devil activity at a given location. The dust devils are imaged by cameras on Pathfinder lander, Spirit rover, Opportunity rover, Phoenix lander and Curiosity rover [130, 131, 133–136, 215]. The Viking Landers 1 and 2 did not image any dust devils on Mars [216], whereas Opportunity and Curiosity rovers have imaged a limited number of active dust devils to date. The Pathfinder and Phoenix were short term missions due to which the images recorded by them could not tell about the

seasonal pattern of dust devil activity on Mars over years. The Spirit rover housed the longest-lived surface-based camera that imaged enough dust devils to study the seasonal as well as diurnal variation of dust devils. It has imaged dust devils for three Martian years in Gusev crater ($\sim 14^{\circ}$ S, $\sim 175^{\circ}$ E), indicating that dust devil activity increases as the local summer approaches in each hemisphere [134]. The diurnal variation of dust devil activity at the Spirit site peaks during the early afternoon [134].

Apart from cameras, some of the landers/rovers discussed above, also carry meteorological suites which record the meteorological parameters of the atmosphere (like pressure, temperature, wind speed, Ultra-Violet (UV) intensity), and can help to detect convective vortices and dust devils. In what follows, we shall now discuss this method of vortex detection.

3.2 Observation of dust devils using meteorological parameters

The convective vortices are detected by distinctive signatures in meteorological parameters as recorded by surface-based instruments. A pressure drop within a convective vortex is generally accompanied by a rise in air temperature and a change in wind velocity [50, 143, 217]. Hence short-term changes in these fields over periods of a few seconds to a minute, suggest the presence of convective vortices. These signatures have been recorded in the meteorological data of Viking Landers 1 and 2, Pathfinder lander, Phoenix lander, and Curiosity rover [23, 26, 135, 137, 140, 142, 218]. If the convective vortex occurrences are also associated with a dip in the UV flux

recorded at the surface, then it suggests the presence of a dust devil [137, 219, 220]. The Viking Landers provided the first opportunity to detect convective vortices using meteorological signatures on Mars. The Viking Landers' wind observation (consisting of both magnitude and direction) was used to detect the presence of a convective vortex, which was occasionally accompanied by atmospheric temperature increase [142, 218]. Subsequently, pressure data from Pathfinder, Phoenix, and Curiosity missions were used to detect convective vortices.

The Pathfinder mission lasted for 83 sols in the valley, Ares Vallis (~19° N, ~33° W). Murphy and Nelli, (2002) [140] identified 79 vortices with pressure drop magnitudes equal to or exceeding 0.5 Pa, from the Pathfinder pressure dataset. The maximum pressure drop magnitude, amongst the vortices identified, was 4.8 Pa at 11:32 Local True Solar Time (LTST). Due to discontinuous temporal coverage of the meteorological data, the actual number of detectable vortices were estimated to be 210 occurring during the 83 sol mission, i.e, ~2.5 vortices per sol [140]. The diurnal variability of these vortex occurrences shows peak during the early afternoon. Due to operational or calibration issues of the Pathfinder wind sensor, no systematic study of the vortex winds could be done, but the wind sensor did provide signals that were qualitatively correlated to dust devil occurrences. Figure 8 in Schofield et al., (1997) [23] shows a substantial and abrupt change in wind direction and speed at the time of measured dust devil pressure signature.

The Phoenix landed in Green Valley (~68° N, ~126° W) with a mission duration of 151 sols. Ellehoj et al., (2010) [135] identified 502 vortices from the entire pressure dataset having pressure drop magnitudes equal to or exceeding 0.3 Pa (~3.3 vortices

per sol), with 197 occurrences having a magnitude greater than 0.5 Pa (detection threshold used by Murphy and Nelli, (2002) [140]). The maximum pressure drop magnitude was 3.6 Pa at 15:08 Local Mean Solar Time (LMST). The diurnal variability of these vortex events shows a peak around noon, just as Pathfinder mission found. The Phoenix lander operated for a comparatively longer duration than the Pathfinder, and hence could detect seasonal variation in vortex activity. The number of identified pressure drops generally increased around Phoenix mission sol 75 (Ls ~111°), which is about 40 sols after the summer solstice. Moreover, the events with large pressure drops increased during the same time [135].

After Pathfinder and Phoenix landers, another rover named Curiosity recorded meteorological data, using which convective vortices are identified. The Curiosity rover landed on Mars in August 2012, and is still operational, thus letting us study the diurnal as well as seasonal dependence of the vortex activity. In the next section, we provide details regarding the mission and dataset used in this study. Next, we provide details of the method used for identifying vortices from pressure excursions. We then discuss implementation of a power-law function to describe the pressure drop statistics. This allows our study to extrapolate for more extreme events and understand whether the detection efficiency has reduced towards the smaller end. Next, we examine other physical parameters such as air temperature and UV radiation flux during pressure drop events, to identify dust devils and the dust loading in the atmosphere due to these events. We also study the seasonal variation of UV drops and the dust devil activity (DDA), which is determined from Mars WRF model. We find a significantly higher frequency of convective vortices that are associated with UV drops, indicating that they contained dust, compared with previous studies [135, 137, 221, 222]. Finally, we use the Burgers vortex model [184] to estimate the minimum tangential velocity of the wind across these vortices.

3.3 Curiosity mission and data description

In this study, we use data from the Rover Environmental Monitoring Station (REMS) instrument on-board the Mars Science Laboratory (MSL) rover Curiosity [223] to identify vortices and estimate the tangential wind velocity of these vortices. The Curiosity rover landed in Gale crater (4.5°S, 137.4°E) on 6th August 2012 (UTC) at areocentric solar longitude (Ls) ~151° during the late Martian southern hemisphere winter [26, 224]. The primary goals of this mission are to determine the landing site's habitability by landing in a place with past evidence of water, characterize the climate of Mars, characterize the geology of Mars, and prepare for a future manned mission to Mars. The Curiosity rover carries a suite of 10 instruments, which includes cameras, spectrometers, radiation detectors, environmental sensors, and atmospheric sensors [225].

The REMS instrument on-board MSL measures air temperature, ground temperature, atmospheric pressure, wind speed and direction, atmospheric relative humidity, and UV radiation fluxes [223]. The baseline strategy for REMS operation is recording of data for every 5 minutes at each Martian hour, every sol, at 1 Hz for all sensors. Additional 1 h "extended blocks" are added into every sol in order to produce complete diurnal coverage every 6 sols. A high solar irradiance leads to the heating of

the surface, which is considered to be the primary reason for the generation of daytime convective vortices. In this study, we focus on the daytime convective vortices; therefore, we only utilize the data from times when solar irradiance is highest on Mars. We study data between 8:00 to 17:00 LTST for each mission sol ranging from 1019 to 1686 (MY 33), which is publicly available in the NASA Planetary Data System [226, 227].

The REMS wind sensor on boom 1 was found to be partly damaged after MSL landing [228]. The data recorded by the other wind sensor on boom 2 was reliable for winds coming from the hemisphere in front of the rover [228], which is achieved for only a portion of each sol. Many efforts of recalibration of the wind measurements were initiated, but they include only median horizontal wind velocities binned every 5 Martian minutes [137]. Since we require data with a timescale of seconds, we could not utilize the available calibrated wind data in this study. In addition, the wind sensor on boom 2 ceased working 2.4 Mars years into the mission, hence no wind data are available from this point onward [229].

We also utilize atmospheric aerosol optical depth (880-nm) retrieved using Mast camera [230] for mission sols 1545 – 1660. Mastcam images of the Sun are nominally taken every three to seven sols [231]. By looking directly at the Sun with Mastcam, the amount of energy reaching the surface can be determined. The aerosol optical depth is derived from direct imaging of the Sun and line-of-sight (LOS) extinction by Mastcam [136, 232–235].

3.4 Methodology to identify vortices and determine Dust Devil Activity

Thermally driven convection is a significant driver for vortex and dust devil generation in the atmospheric boundary layer [9]. As already described in section 3.2, the main signatures of a convective vortex are a drop in pressure, a rise in temperature, and changing wind speed and direction [137, 148]. A temperature variation is difficult to measure since it requires a temperature sensor with a fast response time. Also, the previous Mars lander Phoenix observed frequent short duration pressure drops, however, increases in temperature only occasionally accompanied them [135]. In the case of MSL, calibrated wind data with the timescale of seconds are not available. Hence, we identify convective vortices by primarily searching for transient pressure drops.

The pressure drop dataset used in this work is that obtained by Newman et al., (2019) [41]. Their pressure drop algorithm uses a running-average treatment with three intervals of 20 seconds each, similar to that of Ellehoj et al., (2010) [135] and Kahanpaa et al., (2016) [137]. The algorithm searches for a 20-second interval, which fulfills the following criteria: (a) Minimum pressure is more than 0.5 Pa lower than the average of the previous and following 20 s intervals. (b) Minimum pressure is more than 0.3 Pa lower, and mean pressure more than 0.1 Pa lower, than the mean pressure in the previous and following 20 s intervals.

The minimum threshold for detection of vortices was taken to be 0.5 Pa since the REMS pressure sensor has a peak-to-peak noise of 0.2 Pa [221, 236]. The first criterion identifies relatively strong pressure drops with a very short duration, which appeared to be prevalent during the second and third Martian years of the MSL mission. The second criterion enables the detection of pressure drops with durations longer than 60s, which is an advantage compared to the methods applied by Steakley and Murphy, (2016) [221] and Ordonez-Exteberria et al., (2018) [237]. There is also an instrumental issue described in Harri et al., (2014) [236] as a "shadow effect" which occurs when the REMS UV sensor is temporarily shadowed by a rover structure [221]. This shadowing leads to a decrease in measured pressure with a magnitude of less than 1 Pa within 2 - 3 min of the start of the shadow occurrence. The events associated with the "shadow effect" which have pressure drops less than 0.8 Pa were omitted for detection of the total number of convective pressure drops. The pressure drop events are fitted with a modeled pressure profile using a linear combination of a Lorentzian function (representing the pressure drop) and a line (representing the slow background pressure trend at that time of sol):

$$p(t) = p_{\infty} + k(t - t_o) - \frac{\Delta p}{\left(\frac{t - t_o}{\Gamma/2}\right)^2 + 1} \qquad \dots (3.1)$$

where, p(t) is the observed pressure as a function of time t, Δp is the pressure drop, p_{∞} is the background pressure at the time of the pressure drop event, t_o is the time of the center of the maximum pressure drop, Γ is the full width at half maximum (FWHM) and k is the slope of the background pressure.

The Lorentzian function is a good approximation of the Burgers vortex model [137, 238]. The Burgers vortex model is a theoretical model for any vortex where the

vorticity is provided by continuous convection-driven vortex stretching [184]. A detailed information about the Burgers vortex is provided in chapter 2 section 2.2.2. Pressure drop events with more than one clear minimum were fitted with a linear combination of two Lorentzian functions. After fitting the events, the events with Δp magnitude smaller than 0.6 Pa were excluded from the study since the pressure data of MY 33 are noisier as compared to the previous year's data, thus making the identification of vortices smaller than 0.6 Pa difficult [41]. We use these fitted pressure drop values to estimate tangential velocities of the vortices at the sensor location using the Burgers vortex theory [184]. We use the following equation to estimate the lower limits of tangential wind velocities of the detected vortices at the location of the pressure sensor [51, 179, 184]:

$$\Delta p = -\frac{\rho V^2}{2} \qquad \dots (3.2)$$

where, *V* is the tangential velocity of the vortex and ρ is the density of ambient air. The value of ρ is calculated at the time of the detected pressure drop events using the equation $\rho = \frac{p}{R.T}$, where T is the air temperature measured by REMS, *p* is the pressure, and R is the gas constant of the air in the atmosphere (R=192 m²s⁻²K⁻¹ for Mars [9, 239]). Equation (3.2) suggests that the tangential velocity of wind around a vortex is an explicit function of the pressure drop across the vortex and not its size and varies as the square root of the pressure drop magnitude.

A drop in the UV flux at the time of a pressure drop event indicates atmospheric dust loading caused by the vortex. We utilize the UV flux observations by REMS to detect possible dust lifting in the convective vortices. The UV drops corresponding to each vortex event are identified by visual inspection, and only consider UV drops with a magnitude greater than 0.2%.

The Dust Devil Activity (DDA) has been used to compare vortex observations to dust devils [41, 137]. The DDA is defined as the flux of energy available to drive the dust devils in the atmosphere. It is given as [143],

$$DDA \approx \eta F_s$$
 ...(3.3)

where, η is the vertical thermodynamic efficiency of the dust devil and F_s is the surface sensible heat flux. A detailed information about the DDA is provided in chapter 1 section 1.5.2.2. We calculate the DDA at MSL's location over Martian years 32 - 34using output from the Mars WRF atmospheric model. Mars WRF is the Weather Research and Forecasting (WRF) model for Mars based upon the terrestrial WRF model developed by the National Center for Atmospheric Research (NCAR) [41, 229, 240– 243]. In this work we use similar model configuration and setup as described in Newman et al., (2019) [41].

The Mars WRF is run as a global 2° model with 4 "nested" higher resolution regions (5 domains) roughly centered on MSL's landing site, with each nest smaller than its parent and with higher horizontal resolution. We use results from vertical grid B as described in Newman et al., (2017) [229], as it produces the best match to winds and Aeolian features within Gale Crater [41]. The output from the Mars WRF simulation is generated for the time-varying, three-dimensional atmospheric dust distribution as prescribed by the Mars Climate Database (MCD) MGS dust scenario for

a year without any major dust storm [244, 245]. The output is generated for 12 simulations, each lasting for eight sols at solar longitude, Ls = 0, 30, 60, 90, 120, 150, 180, 210, 240, 270, 300, and 330°, to fully sample the annual cycle of solar forcing. This output is then interpolated in time to every sol of the MSL mission, and also interpolated spatially to the rover's location in that sol. The spatially varying surface properties are taken from MGS Mars Orbiter Laser Altimeter (MOLA) and Thermal Emission Spectrometer (TES) observations [246]. These properties are MOLA topographic height [247], MOLA surface roughness map [248], and albedo, thermal inertia, and emissivity maps [249]. Moreover, high-resolution topography data are used for domain 5 which is taken from a Mars Express, High Resolution Stereo Camera Instrument Digital Terrain Map [250, 251].

The PBL top, which is needed to find p_{top} in equation (1.9) as mentioned in chapter 1, is calculated inside Mars WRF's boundary layer scheme as described in Hong and Pan, (1996) [252]. The sensible heat flux is also computed inside Mars WRF surface layer scheme [253], which uses Monin-Obukhov similarity and accounts for four stability categories: stable, mechanically induced turbulence, unstable forced convection, and unstable free convection [41]. The DDA has also been used to parametrize dust devil lifting in Mars atmospheric models [46].

3.5 Results and discussions

We analyze here REMS data for MY 33 (mission sols 1019 to 1686). However, in the latter part of this section, we also compare MY 33 with other Martian years from previously published results and with Mars WRF simulations.

3.5.1 Identification of vortices and diurnal variation

Figure 3.2 shows the numbers of identified pressure drop events per sol during MSL mission sols 1019 to 1686 for $\Delta p \ge 0.6$ Pa, $\Delta p \ge 1.5$ Pa, and $\Delta p \ge 3.0$ Pa distributions, which are divided as per seasons on Mars. Out of a total of 668 sols studied, vortices only occur during 299 sols. Figure 3.2 indicates that the frequency of vortex formation and its detection increases as the season advances into the southern hemisphere summer (Ls = 270° to 360°).



Figure 3.2: The numbers of identified pressure drop events for each sol of MY33 during MSL mission are shown for $\Delta p \ge 0.6$ Pa, $\Delta p \ge 1.5$ Pa and $\Delta p \ge 3.0$ Pa distributions. Each plot corresponds to four seasons on Mars.

One of the reasons for increased vortex activity is the higher solar insolation received during the summertime, which enhances the convective processes. Another reason could be the position of the rover. As discussed in section 3.4 (equation (3.3)), the DDA depends on two quantities, the thermodynamic efficiency of the vortex and the sensible heat flux of the surface. The sensible heat flux depends on the temperature difference between the surface and air, near-surface air density, winds in the region driven by the topography, and the thermal inertia of the surface. Whereas, the thermodynamic efficiency depends on the pressure thickness of the PBL. The sensible heat flux generally peaks around noon, whereas the thermodynamic efficiency peaks between ~14:00 and ~16:00 LTST, shifting later as the season shifts from spring to summer to fall. This signifies that the peak DDA generally occurs around the local summer season, depending mainly on a stronger afternoon sensible heat flux. Around local winter solstice, the surface-to-air temperature difference and PBL depth are both far lower than in any other season, which results in the lowest predicted DDA during the winter season. The DDA remains intermediate during the spring and autumn local seasons. The sensible heat flux has increased each year as the rover climbs up the slopes of Aeolis Mons, thus increasing the predicted DDA over 3 Martian years during the local summer season [41], as will be seen later in this section in Figure 3.7.

In the southern summer season, we observe a significant increase in the frequency of strong vortices ($\Delta p \ge 1.5$ Pa) during afternoon hours (between 11:00 and 14:00 LTST), as also reported in Figure 9 of Newman et al., (2019) [41]. The enhancement is mainly caused by heating of the surface by solar radiation around noon [48, 142, 254–256]. The occurrence of vortex activity diminishes sharply after 16:00 LMST, which can be related to the collapse of the daytime boundary layer [135, 257–260]. The diurnal distribution of vortices detected by Mars Exploration Rover (MER)

Spirit, Mars Pathfinder and MSL REMS also shows that these generally occur during 11:00 – 14:00 LTST [133, 134, 137, 140, 221, 237].

Figure 3.3 shows the cumulative magnitude distribution of the pressure drops (pressure drop events larger than 0.6 Pa) detected in MY33. We apply the power-law fit [219, 220, 237, 261] to the cumulative distribution of the vortices. This gives an insight into the relative occurrence of vortices with respect to their magnitude of pressure drops.



Figure 3.3: The cumulative magnitude distribution of the pressure drops and the corresponding power-law fit slope detected in MY33. The y-axis shows the cumulative number of pressure drop events larger than 0.6 Pa, divided by the total number of sols of vortex observations.

The magnitude of the power-law fit in our study is -2.6. Using REMS data, Steakly and Murphy, (2016) [221], Kahanpää et al., (2016) [137] and Ordonez-Exteberria et al.,

(2018) [237] obtained a power-law fit of -2.77, -2.76 and -3.14 respectively. The powerlaw fits for Pathfinder and Phoenix missions are -1.75 and -2.36 [237], & -1.73 and -2.48 [221] respectively. Kahanpää et al., (2016) [137] reported a power-law fit of -1.68 for Pathfinder data. Generally, a higher value of power-law slope means that the vortices with higher pressure drop magnitudes will be less frequently formed as compared to their weaker counterparts. The power-law slope is lowest for the pressure drop data obtained from the Pathfinder mission, thus meaning that there is a high abundance of stronger vortices at that site. Alternatively, a higher value of power-law slope for MSL data corresponds to a low abundance of stronger pressure drop events. This can be related to the peak PBL heights in Gale crater (most often \sim 3 – 6 km) [41], being lower than those reached at the Pathfinder landing site (upto about \sim 9 km) [262]. The peak PBL heights at the Phoenix landing site are also around \sim 4 km [263].

The differences in the power-law slopes of Pathfinder and Phoenix data in different studies may be caused by a difference in the normalization of vortices per sol, and differences in the weighting of the low-frequency data corresponding to the more extreme events [237]. The difference in power-law slopes for REMS data between Ordonez-Exteberria et al., (2018) [237] and Kahanpää et al., (2016) [137] comes from differences in the searching algorithms for pressure drop events, or the period of analysis. Ordonez-Exteberria et al., (2018) [237] analyzed two Martian years' data instead of only one and identified the rare events with large pressure drops in the second Martian year. Among the power-law slopes for all Martian years at MSL, the power-law slope is least for MY 33. This is consistent with the result of Newman et al., (2019) [41], which shows that during the third operating year of MSL, the number of detected

pressure drops increased along with the DDA. Amongst the MSL observations for the period MY 31 - MY 33, the highest pressure drop event is ~5.6 Pa in MY 33.

Figure 3.4 shows the distribution of event duration (Γ) with pressure drop (Δp). The full width at half maximum (FWHM) duration (Γ) represents approximately half of the total event duration. For MY 33, the mean FWHM of single peak events is 13.7 s as compared to 12.6 s in MY 32. In MY 33, the longest pressure drop event has Γ = 63.6 s and Δp = 1.2 Pa, and the largest pressure drop event has Δp = 5.59 Pa and Γ = 2.7 s. The pressure drop of 5.59 Pa is quite high as compared to the highest pressure drop of ~3.0 Pa reported by Kahanpää et al., (2016) [137]. This is because during MY 33, the rover moved into a region of stronger vortex activity compared to the period analyzed by Kahanpää et al., (2016) [137] (MY 31 – MY 32).



Figure 3.4: Pressure drop magnitude (Δp) versus full width at half maximum, FWHM, (Γ) of all the detected events.

The higher FWHM is due to the slight change in the method of vortex detection, as discussed earlier (section 3.4). Most events have an FWHM duration Γ between 5 and 30 s. The events with larger pressure drop (Δ p) have smaller FWHM durations (Γ) and vice versa. The distance between the vortex and the sensor also plays an essential role in determining the FWHM for a pressure drop event [135, 137]. As the distance between the sensor and the vortex increases, the detected pressure drop magnitude decreases. At the same time, the FWHM duration increases as the shape of the detected pressure drop curve smooths out. For MY 33, we have detected higher number of vortices that have larger pressure drop and smaller FWHM as compared to (MY 31 – MY 32), indicating that a higher number of vortices would have passed relatively near to the sensor.

3.5.2 Variation of temperature and UV index

Figure 3.5 shows the variation of pressure and air temperature with time during a vortex event, 90 seconds before and after the pressure drop event occurred (at Time = 0 s). The event occurred on MSL sol 1546 at 14:37 LTST.


Figure 3.5: REMS measurement of pressure and air temperature during a vortex event on mission sol 1546 at 14:37 LTST.

Here, we observe a simultaneous rise in air temperature along with a fall in pressure. Events like this might happen because the pressure drops of convective vortices should be generally accompanied by a rise in air temperature. However, we observe this trend in approximately 15% of the vortex events only, which could be attributed to several reasons such as: the vortex core is far from the sensor and the air near the sensor is not heated enough due to poor air conductivity; turbulent temperature fluctuations which overcome the temperature perturbation in the vortex [135]; the hot core of the vortex missing the sensor in the vertical direction [135]; vertical flow distortions due to rover structures [135]; and thermal contamination from the rover's energy sources [137].

The REMS instrument records variation in UV intensity in multiple wavelength bands: A (315 - 370 nm), B (280 - 320 nm), C (220 - 280 nm), D (230 - 290 nm), E (300 - 350 nm), and ABC (200 - 370 nm) [223]. If vortices detected from the pressure

data also lift dust, then the dust may obscure sunlight reaching the UV sensors. The detection of a drop in UV intensity is dependent on the fact that the dust-laden vortex must pass between the sensor and the direction of the Sun [219]. This obscuration of sunlight will cause a drop in the observed solar UV intensity. Figure 3.6 shows an obscuration of approximately 3.5% of the mean intensity during one such event ($\Delta p = 3.66$ Pa).



Figure 3.6: REMS measurement of ultraviolet radiation intensities during a vortex event on mission sol 1546 at 14:37 LTST.

A significant drop in UV intensity data indicates that the vortex lifts a larger amount of dust. Out of the 611 detected pressure drops, about 93 events also show a simultaneous drop in UV intensity with a dimming greater than 0.2%. Ordonez-Exteberria et al., (2020) [222] detected only 13 simultaneous UV attenuations in MY 33 as compared to 93 in this study. Apart from different detection algorithms, OrdonezExteberria et al., (2020) [222] only consider those UV attenuations that occur at high sun elevation ($77\pm5^{\circ}$). However, we consider UV attenuation at all sun elevation angles ranging between 8:00 and 17:00 LTST (day duration considered in our study for vortices detection).

Table 3.1 lists the events detected with nearly simultaneous pressure drops and corresponding UV attenuations. We also calculate the optical depth for each of these events using Beer Lambert's Law ($\tau = ln(UV_{background}/UV_{attenuation})$) and is listed in table 3.1.

Table 3.1: List of detected pressure drops with simultaneous UV drop in REMS data. The last column lists the estimated optical depth using Beer Lambert's Law for the respective event.

Mission sol	Ls	Pressure drop (Pa)	UV background (W/m ²)	UV during attenuation (W/m ²)	Percentage drop in UV flux (%)	Optical depth of the vortex event (\tau)
1023	2.33	0.99	11.74	11.72	0.2	0.0017
1039	10.4	0.91	12.37	12.3	0.6	0.0057
1039	10.4	0.61	12.6	12.49	0.9	0.0088
1040	10.89	0.6	12.35	12.31	0.3	0.0032
1040	10.89	0.82	12.37	12.24	1.1	0.0106
1055	18.25	1.03	13.15	13.12	0.2	0.0023
1065	23.08	0.61	13.08	13.01	0.5	0.0054
1066	23.55	0.69	13.15	13.1	0.4	0.0038
1086	33.01	1.87	12.76	12.6	1.3	0.0126

1113	45.51	0.75	10.4	10.34	0.6	0.0058
1176	73.92	1.09	7.98	7.9	1	0.0101
1209	88.78	0.64	7.65	7.59	0.8	0.0079
1335	149.46	0.68	13.55	13.49	0.4	0.0044
1335	149.46	0.6	13.74	13.66	0.6	0.0058
1387	178.03	1.59	14.7	14.54	1.1	0.0109
1400	185.61	1	14.43	14.37	0.4	0.0042
1404	188.01	1.32	15.94	15.9	0.3	0.0025
1405	188.58	1.21	15.24	15.2	0.3	0.0026
1406	189.16	1.11	13.66	13.59	0.5	0.0051
1410	191.56	0.61	15.37	14.94	2.8	0.0284
1417	195.77	4.24	12.48	11.57	7.3	0.0757
1421	198.21	0.61	14.82	14.77	0.3	0.0034
1431	204.33	0.89	15.07	15.02	0.3	0.0033
1449	215.57	0.74	13.51	13.48	0.2	0.0022
1455	219.38	1.2	15.33	15.15	1.2	0.0118
1459	221.91	1.13	12.97	12.92	0.4	0.0039
1493	243.9	0.96	15	14.73	1.8	0.0182
1493	243.91	1.34	15.1	14.8	2	0.0201
1494	244.55	5.59	15.25	14.68	3.7	0.0381
1500	248.45	0.88	15.97	15.94	0.2	0.0019
1512	256.28	0.91	15.81	15.76	0.3	0.0032
1516	258.84	1.24	7.35	7.32	0.4	0.0041
1518	260.17	0.68	9.87	9.83	0.4	0.0041
1522	262.85	1.66	12.01	11.95	0.5	0.005
1522	262.87	2.69	7.42	7.28	1.9	0.019
1527	266.11	0.69	8.45	8.34	1.3	0.0131
1534	270.61	1.99	13.57	13.46	0.8	0.0081
1539	273.75	3.72	5	4.89	2.2	0.0222
1539	273.86	1.13	9.15	9.01	1.5	0.0154
1541	275.1	0.71	13.87	13.79	0.6	0.0058
1544	277.06	1.16	13.38	13.33	0.4	0.0037
1546	278.38	3.66	6.15	5.92	3.7	0.0381

1554	283.38	1.19	17.8	17.66	0.8	0.0079
1555	284.08	3.6	17.91	17.39	2.9	0.0295
1556	284.69	1.16	19.52	19.41	0.6	0.0057
1562	288.44	0.83	17.63	17.57	0.3	0.0034
1562	288.46	0.69	19.98	19.93	0.3	0.0025
1565	290.35	2.77	19.68	19	3.5	0.0352
1565	290.36	0.99	20.18	20.1	0.4	0.004
1565	290.36	0.99	20.27	19.98	1.4	0.0144
1565	290.37	0.61	20.09	20.05	0.2	0.002
1566	291	1.63	20.73	20.55	0.9	0.0087
1567	291.57	0.69	14.12	14.05	0.5	0.005
1567	291.57	3.36	14.74	14.12	4.2	0.043
1569	292.91	1.94	13.07	12.98	0.7	0.0069
1569	292.92	0.86	9.5	9.33	1.8	0.0181
1570	293.49	5.57	20.33	19.42	4.5	0.0458
1570	293.51	0.71	20.1	20.01	0.4	0.0045
1570	293.52	2.61	18.18	18.02	0.9	0.0088
1570	293.54	1.97	11.29	11.06	2	0.0206
1578	298.41	1.66	10.56	10.4	1.5	0.0153
1578	298.42	2.38	12.27	12.04	1.9	0.0189
1581	300.33	0.62	14.44	14.35	0.6	0.0063
1581	300.35	0.81	10.3	10.23	0.7	0.0068
1587	303.96	1.3	18.16	17.9	1.4	0.0144
1590	305.77	1.51	16.29	16.1	1.2	0.0117
1594	308.24	0.64	17.34	17.16	1	0.0104
1595	308.8	1.67	18.55	18.4	0.8	0.0081
1596	309.42	3.63	18.43	18.17	1.4	0.0142
1604	314.18	0.81	16.98	16.88	0.6	0.0059
1605	314.79	3.54	15.28	14.67	4	0.0407
1609	317.14	0.65	17.65	17.52	0.7	0.0074
1618	322.48	1.16	4.76	4.71	1.1	0.0106
1620	323.62	0.68	8.14	8.05	1.1	0.0111
1629	328.69	1.78	13.32	13.03	2.2	0.022

1630	329.27	1.23	12.6	12.31	2.3	0.0233
1635	332.05	1.68	12.31	12.11	1.6	0.0164
1636	332.66	1.27	13.72	13.59	0.9	0.0095
1637	333.19	0.63	12.93	12.87	0.5	0.0047
1641	335.49	0.91	9.72	9.6	1.2	0.0124
1642	335.99	0.62	13.52	13.39	1	0.0097
1645	337.68	0.73	12.6	12.44	1.3	0.0128
1645	337.69	2.71	11.65	11.5	1.3	0.013
1646	338.22	1.15	12.49	12.39	0.8	0.008
1650	340.4	1.64	12.36	12.13	1.9	0.0188
1658	344.77	0.76	12.19	12.16	0.2	0.0025
1660	345.8	1.32	13.51	13.45	0.4	0.0045
1666	349	0.61	14.31	14.21	0.7	0.007
1673	352.74	0.63	12.97	12.9	0.5	0.0054
1680	356.37	0.72	14.04	13.99	0.4	0.0036
1680	356.38	0.68	13.5	13.42	0.6	0.0059
1682	357.41	0.64	13.91	13.87	0.3	0.0029
1686	359.48	0.62	14.17	14.1	0.5	0.005

Figure 3.7 shows the seasonal variation (with Ls) of the UV attenuation events and the seasonal variation of the dust devil activity (DDA) for MY 32, MY 33 and MY 34. The DDA are predicted by the Mars WRF model using the same method and model setup as described in Newman et al., (2019) [41] and as discussed in section 3.4 above. A prescribed dust scenario was used to produce these results, thus any year-to-year changes at the same Ls are purely due to the changing position of the rover.



Figure 3.7: (a) Variation of events with nearly simultaneous UV attenuation with Ls. The dotted green lines indicate the season boundaries. Different seasons indicated are for the southern hemisphere of Mars. (b) Variation of dust devil activity (DDA) with Ls for three Martian years.

We observe that the number of UV drops detected is the least during local winter season. This similar trend is seen in DDA, which is minimum (~0.08 for MY 33) during the same season. Generally, the number of events (and their strength) with UV attenuations increases as we move towards the local summer season (Figure 3.7 (a)). This may be due to either stronger vortex forming at this time of year, resulting in more raising of dust, or more loose dust being available to be lifted at the rover's location at this time of year. The former is likelier, however, given that – as found in e.g. Newman et al., (2019) [41] – the peak daily DDA is predicted to be largest during a 90° L_s period centered on local summer solstice in every year, with the preceding 90° L_s period (centered on southern spring equinox) being a close second. In that work, Mars WRF

output is used to attribute this to a combination of a much higher PBL height and slightly stronger sensible heat fluxes in those periods, compared to other times of year (Newman et al., (2019) [41], Figure 10). We observe a significant outlier event with a 7.3% UV attenuation event during local spring (Ls = 195.77°), which could be caused due to various local factors such as local topography, surface thermal conductivity, and availability of dust.

During the local summer season, as the number of events and intensity of UV drops increases, the DDA shows a corresponding increase, as compared to other seasons (Figure 3.7(b)). DDA represents the vortex activity, but cannot distinguish whether a vortex is dusty or not. The stronger DDA relates to stronger/greater numbers of vortices and so might relate to them being able to lift more dust. The DDA for MY 32 is least as compared to MY 33 and MY 34. The highest DDA during the southern hemisphere summer of MY 33 corroborates with the detection of a high number of dust devils (events with UV, as well as pressure, drops) in this study. There are many vortices with high-pressure drops that do not show a simultaneous UV drop. However, this does not necessarily prove that such events were not capable of lifting dust, as it's possible that they did not pass between the rover and Sun to cast any shadow on the sensor.

We estimate the optical depth ranging from 0.0017 to 0.0757 by directly utilizing the UV attenuation magnitude. Reiss et al., (2014) [211], utilized "shadow method" to calculate the optical depth of three dust devils imaged by HiRISE, and reported the optical depth ranging from 0.29 ± 0.18 to 1.20 ± 0.38 . The lower estimation of optical depth in this study could be attributed to three major causes: 1) The dust

devils were far away from the rover at the time of detection which causes a lower UV attenuation; 2) The location of the dust devils observed by HiRISE (42° S, 108.6° E; 68.6° S, 11.4° E and 60.8° N, 212.3° E) were distant from the Gale crater (3.4° S and 137° E), and thus we can expect a difference in the amount of surface dust available for lifting; 3) The difference in size of the devils, as there is an obvious bias in that only the largest dust devils will be observed from orbit. The total background atmospheric optical depths using observations by Mastcam (880 nm wavelength band), onboard the Curiosity rover, peak during the local summer season in MY 32 and have been reported to be in the range ~0.6 to ~1.4 [137, 232]. We have retrieved the optical depths from Mastcam, for the southern summer season of MY 33, and found values similar to those reported for MY 32.

Out of the 611 pressure drop events, nearly ~86% of the events do not show any UV attenuation, whereas only ~5% events gave attenuation greater than 1%, and none of the attenuations crossed 10% (Figure 3.8). The ranked UV attenuation of Ordonez-Exteberria et al., (2020) [222] indicated that only ~4% of events show any kind of measurable attenuation. Comparing it to our Earth counterpart, we see that nearly ~60% of the events on Earth show a measurable attenuation with around ~40% showing attenuation higher than 1%, and about ~10% with attenuation higher than 10% [219]. This indicates that the vortices on Earth in general can lift more dust compared to Martian vortices, which could relate to the difference of the dust-size between the two planets.



Figure 3.8: Ranked UV attenuation of the 611 pressure drop events showing ~86% events has no measurable UV attenuation.

An in situ sampling of vertical size distribution of dust grains in a terrestrial dust devil shows that ~80 μ m sized particles can be found near the surface, while tens of micron sized particles are found to be lifted inside the dust devil [264]. Such in situ measurements inside Martian vortices are not available, but remote sensing observations show that majority of dust particles on the surface of Mars are of micron size [78]. It has also been estimated that the size of dust particles entrained in the atmosphere, correspond to only a few microns [265]. Dust devils can lift particles into the atmosphere either due to the frictional drag of vortex winds moving over a sand bed, or the due to the low pressure at the center of the vortex (the so called Δp or suction effect). It has been shown that both these thresholds are higher for smaller particles [149]. Bila et al., (2020) [266] performed an experiment to study dust lifting due to suction effect within a dust devil in Martian condition, and showed that high pressures are required to lift more dust particles. Thus the larger sized dust particles on Earth favor the vortices to be more dust lifting compared to Mars.

Figure 3.9 shows the variation of UV attenuation with the measured pressure drop. The red line in the plot is an envelope of the UV attenuation in the observed data, given by the function A (%) = $3 \times \Delta p$ (Pa).



Figure 3.9: Scatterplot of UV attenuation and measured pressure drop. The red line is an envelope of the data given by the function $A(\%) = 3 \times \Delta p$ (Pa).

The envelope factor of 3, is much less than the corresponding factor of 50 for terrestrial dust devils [219]. This shows that the maximum attenuation expected for a given pressure drop is much less than expected for the same pressure drop in dust devils on Earth. We do not observe any significant UV attenuation associated with weak pressure drops, which is also evident with terrestrial dust devils [219]. A recent

laboratory study [266] on particle lifting in Mars-like conditions shows that the first grains start to lift at only 2.0 ± 0.8 Pa, and more grains will be lifted as we increase the pressure drop to 10 Pa. Therefore, any vortex must have a strong pressure drop at the core to eventually become a dust devil.

The detection of 611 vortices in 669 sols in the present study and that in Newman et al., (2019) [41], translates into a ~91% frequency of occurrence of vortices per sol. In the present study, we further identified ~15% of these as likely to be dust devils. On the other hand, Kahanpää et al., (2016) [137] and Steakley and Murphy, (2016) [221] obtained convective vortex occurrence frequencies of ~38% and ~35% respectively for the duration, MY 31 ($L_s=158$) – MY 32 ($L_s=158$). Newman et al., (2019) [41] have reported the frequency of occurrence of ~45% in MY 32.

3.5.3 Physical parameters related to vortices

Figure 3.10 shows the variation of estimated tangential velocities (V) with pressure drops (Δp) using Burgers vortex theory. This theory suggests that the tangential velocity of the vortex will vary as the square root of the pressure drop magnitude. This suggests that for an increment in pressure drop values towards the higher pressure drops, the increase in their corresponding tangential velocity magnitude will be less significant than the pressure drop magnitude.



Figure 3.10: Tangential velocities (V) versus pressure drops (Δp) as calculated from Burgers vortex theory.

The red curve (Figure 3.10) shows the square root fit of the tangential velocities with respect to the pressure drop. These tangential velocities are just a lower limit estimation to the corresponding pressure drops. The real tangential velocity can only be determined if a vortex passes directly over the sensor.

Dust devils play an important role in the entrainment of dust in the Martian atmosphere. However, complete information about the method by which fine dust particles of size few microns are lifted from the surface to the atmosphere is not fully understood yet. In the Martian atmosphere, it has been estimated that vortices could have a threshold tangential velocity of $\sim 20 - 30$ m/s [149]. This laboratory experiment used a vortex generator with a 10 mbar pressure and demonstrated lifting of dust of size 2 µm. Assuming the dust devils to be in cyclostrophic balance [143], the pressure drop in the center of a vortex is expressed as,

$$\Delta p = \frac{V^2}{RT} p_{avg} \qquad \dots (3.4)$$

where, R is the gas constant of the air in the atmosphere, T is the mean temperature of the atmosphere, p_{avg} is the mean pressure of the atmosphere, and V is the maximum tangential wind around the vortex. If we assume typical values of T = 250 K, R = 192 m²s⁻²K⁻¹, and p_{avg} = 700 Pa for Mars, we obtain $\Delta p \approx 13$ Pa for V = 30 ms⁻¹.

We do not find any vortex with such high-pressure drops in our analysis. The tangential wind velocities in our study range from ~ $7.0 - 25.0 \text{ ms}^{-1}$, with a maximum velocity of ~ 25.0 ms^{-1} corresponding to the maximum Δp of 5.6 Pa. A total of 93 cases in our study shows a simultaneous decrease in UV flux, indicating dust lifting at the time of vortex occurrence. Our estimated wind velocities are well below the Martian dust lifting threshold, mainly due to two possible causes. The first being that the tangential winds are not the only mechanism sufficient for lifting dust in vortices forming dust devils. The dust devils lift dust by a combination of near-surface wind stress and impact saltation [46, 133]. Another reason would be that we lack knowledge of the distance between the vortex and the instrument. If a vortex did not pass just over or very close to the instrument, then it becomes challenging to know the exact pressure drop inside the vortex. The pressure drop variation with the radial distance from the vortex center is given as [267]:

$$\Delta p(r) = \Delta P \left(1 + \frac{r^2}{r_m^2} \right)^{-1} \qquad \dots (3.5)$$

where, ΔP is the pressure drop in the vortex center, $\Delta p(r)$ is the pressure drop of the vortex at a distance r and r_m is the size of the vortex at which the tangential velocity is at its maximum.

Based on the observations of the dust devils on Mars, the diameter of the dust devils can range from 15 m to 280 m [127, 211]. Figure 3.11 shows the variation of the detected pressure drop as a function of the radial distance between the instrument and the vortex, assuming that the central pressure drop is the 13 Pa required to form a dust devil (the threshold for $V = 30 \text{ ms}^{-1}$). This plot suggests that the larger the size of the vortex, the smaller will be the rate of decrease of pressure drop with radial distance.

For a vortex having a pressure drop of 13 Pa and a radial size of 50 m, the observed pressure drop will become less than the threshold pressure drop of 0.5 Pa after a distance of approximately 250 m. This indicates that the instrument would not be able to detect this vortex if this vortex forms at a distance greater than 250 m from the instrument. Alternatively, if this vortex is in the proximity of within 250 m and is not passing exactly over the instrument, the instrument would record a lower pressure drop as compared to the actual pressure drop within the vortex. This would lead to an incorrect estimation of the maximum tangential wind velocity for the particular vortex.



Figure 3.11: Variation of pressure drop around a vortex as a function of radial distance from the vortex. The black dashed line is the reference for the threshold pressure drop of 0.5 Pa for the detection of the vortex in our study.

3.6 Conclusion

Many previous missions with landers/rovers (two Viking Landers, Mars Pathfinder, Mars Phoenix, Spirit, Opportunity, and Curiosity) have identified convective vortices in the Martian atmosphere [112, 130, 131, 135, 142, 218]. We have analyzed the pressure drop dataset from REMS on the Curiosity rover for MY 33 (mission sols 1019 - 1686) between 08:00 to 17:00 hours (LTST), with a total of 611 daytime convective vortices having pressure drop $\Delta p > 0.5$ Pa [41]. The vortex activity peaks during noon hours due to higher solar insolation, with the largest pressure drop of 5.6 Pa at 11:58 LTST on sol 1494 (Ls = 244.5°). The maximum numbers of convective vortices are identified during the southern hemisphere summer season. The

southern hemisphere receives higher solar insolation during this time, and a variation of the PBL height and the sensible heat flux to a lesser extent, controls the seasonal variations of convective vortices.

Compared to some previous studies covering MY 31 – 32 [135, 137, 221], we observe a significantly higher frequency of convective vortices in MY 33. Our powerlaw fit also indicates the increased activity of convective vortices during the period of study. Most pressure drop events have FWHM duration around 20 s, and their pressure drop range between 0.6 Pa to 5.6 Pa. The longest pressure drop event has $\Gamma = 63.6$ s with $\Delta p = 1.2$ Pa. The largest pressure drop event with $\Delta p = 5.6$ Pa has FWHM of $\Gamma = 2.7$ s.

About 15% of cases also show a simultaneous drop in UV intensity with a dimming greater than 0.2%. The drop in UV flux intensity indicates the presence of dust in the vortices, as dust-laden vortex might obscure the solar UV flux intensity reaching the instrument. Generally, the number of events (and their strength) with UV attenuations increases as we move towards the local summer season. The stronger DDA during local summer, relates to greater numbers of vortices and these being able to lift more dust, resulting in more dust devils. Using ranked UV attenuation index, we observe that nearly ~86% of the events do not show any attenuation, only ~5% of events were capable of attenuation greater than 1%, and none of the attenuations crossed 10%.

We estimate the tangential velocity of the wind around the vortices ranges from $\sim 7.0 - 25.0 \text{ ms}^{-1}$. The estimated tangential wind velocities are well below the Martian dust lifting threshold ($\sim 30 \text{ ms}^{-1}$ [149]). This indicates that either the tangential wind

velocities alone may not suffice to lift dust in vortices (to form dust devils), or the detected vortices are quite far from the observing instrument. The inclusion of other dust lifting mechanisms like impact saltation in atmospheric models would lead to a better understanding of dust devil formation in the future.

Chapter 4 Numerical Simulation of Dust Distribution within a Dust Devil

On Mars there are several observations of active dust devils and their tracks as discussed in chapter 3, indicating that they entrain the surface dust into the atmosphere. A decrease in albedo by $\sim 15\%$ is recorded for regions where surfaces is $\sim 50\%$ covered by dust devil tracks [48, 268]. As also mentioned in chapter 1, dust has a strong impact on the atmosphere's thermal and dynamical state, thus affecting the climate and environment of Mars [52, 57, 138]. Hence, it is important to study about the processes that can entrain dust into the atmosphere and to quantify the subsequent dust flux and loading in the atmosphere. One of the ways by which surface dust can enter into the atmosphere is by "dust devils". But it is difficult to estimate the contribution of dust devils to the total dust loading in the atmosphere. This difficulty is due to the limitation of our knowledge about the amount of dust getting lifted by an independent dust devil, and its frequency of occurrence [210]. In this chapter, we will model the spatial distribution of dust concentration within a steady state Martian dust devil. We will numerically solve the equations of motion for dust particles to determine their velocity inside the dust devil and consequently determine the dust distribution using the continuity equation.

4.1 Introduction

Dust devils are vertical convective vortices with spinning columns of air which are visible by entrained dust and tend to occur on both Earth and Mars [48]. As already described in chapter 1, the Martian atmosphere lacks moisture due to which the Martian surface is dustier than Earth's surface. This eventually leads to the formation of tracks on the Martian ground at the time of passage of a dust devil. On Earth, dust devils are potentially hazardous phenomenon affecting the atmospheric circulations and aerosol transport [269, 270], whereas on Mars they play an important role in dust lifting and thereby affecting the radiative and dynamic properties of the atmosphere [131]. As described in chapter 1, dust devils are formed by heating of near surface air resulting in temperature and pressure deviations, leading to a formation of unstably stratified atmosphere [51]. The dust devil has a low pressure at its interior and is surrounded by high tangential winds [168] and strong vertical velocities. The tangential winds and updrafts maintain vorticity of these structures [46]. The dust devils are very efficient at sucking in available dust from the surface within the convective plume due to the pressure drop generated at the center [46]. This pressure drop (or suction) is intrinsically dependent on the tangential wind velocity around the dust devil (vortex) [135]. The higher the magnitude of tangential wind velocity, the higher will be the pressure drop within the vortex, leading to lifting of more dust [266]. The background dust haze in the Martian atmosphere, during non-dust storm seasons, is believed to be sustained by dust devils [9, 216].

As mentioned in chapter 3, dust devils are often taller than they are wide and vary in shape from columnar to inverted cones to disordered rotating dust clouds [48]. Atmospheric models can determine the dust flux due to convective turbulence in the atmosphere based on the parameterizations for the process. They predict that this daytime convective activity contributes to $\sim 30 - 50$ % of the global Martian dust budget [49, 87]. Apart from the models, space based images have also identified dust devils as a dominant source of dust in the Martian atmosphere [115]. The dust devils are considered to contribute >25 - 75 % of the estimated global dust flux, which is similar to that predicted by the atmospheric models [115, 271]. They transport the dust vertically and later regional winds transport the dust in suspension over long distances for hours or days [272]. However, the quantitative contribution of dust devils to the dust budget of the Martian atmosphere is yet not known clearly. This owes partly to the lack of knowledge on the distribution of dust within the devils, though there are indications that mostly the dust density is highest near the surface [48].

To bridge this gap in our knowledge, we determine the spatial distribution of dust density and the particle velocities within a steady state dust devil. There are several high resolution models to simulate for the dust devil on Mars, but parameterizing their dust lifting abilities along with their temporal and spatial distribution into the Martian global dust cycle is still problematic. For this, we numerically solve equations of motion for dust particles to determine their velocity inside the steady state dust devil. We then estimate the spatial distribution of dust concentration within the devil, by solving the continuity equation. Our calculation domain is an azimuthally symmetric cylindrical coordinate system which is a good approximation for the shape of a real dust devil [168, 195, 273, 274]. Earlier studies on the numerical simulation of dust devils and convective vortices on Mars dealt with the development (dynamical evolution) of these structures in the atmosphere. They predict the track of dust particles within the dust devils, but do not estimate the dust concentration within [51, 190, 195, 275]. Moreover, there are no observations of dust density distribution inside dust devils on Mars, leave alone model estimates. Hence, we first validate our model for the case of dust devils on Earth, where observations are available. Our results indicate that major particle load is near the surface and at the boundary of the dust devil, leading to the formation of sand skirt.

4.2 Model setup and boundary conditions

We assume a dust devil consisting of a steady state convective plume. Development of a convective vortex can be simulated by large eddy simulations (LES) [190, 195, 276] which start with some background condition and with time reach a steady state vortex. Here in our study, we start with a steady state vortex which is able to lift dust. We consider the following net forces acting on dust particles due to vortex winds: drag, lift and gravity. The drag force maintains the forward motion, while the lift force provides the vertical motion to the particles which is counter-balanced by the gravitational force. To maintain suspension of the particles, a higher lift force than gravitational force is required. We consider the following total force acting on a dust particle:

$$F_{p} = \frac{1}{8}\pi\rho C_{D}D^{2}\left|\overrightarrow{U_{r}}\right|^{2} + \frac{1}{8}\pi\rho C_{L}D^{2}\left|\overrightarrow{U_{r}}\right|^{2} - \frac{1}{6}\pi D^{3}(\rho_{p} - \rho)g\hat{z} \qquad \dots (4.1)$$

where, $\overrightarrow{U_r}$ is the relative velocity of the particle with respect to vortex winds, ρ is the density of the air, ρ_p is the particle density, D is the diameter of the particle (spherical), C_D is the drag coefficient, C_L is the lift coefficient and g is the acceleration due to gravity of the planet.

Since, we are considering a cylindrical vortex, we can use the corresponding co-ordinate system, in which case, the time derivative of the particle velocity can be written as following:

$$\frac{d\vec{u'}}{dt} = \left[\frac{\partial u'_r}{\partial t} + u'_r \frac{\partial u'_r}{\partial r} + \frac{u'_{\theta}}{r} \frac{\partial u'_r}{\partial \theta} - \frac{{u'_{\theta}}^2}{r} + {u'_z} \frac{\partial u'_r}{\partial z}\right] \hat{r} + \left[\frac{\partial u'_{\theta}}{\partial t} + u'_r \frac{\partial u'_{\theta}}{\partial r} + \frac{u'_{\theta}}{r} \frac{\partial u'_{\theta}}{\partial \theta} + \frac{u'_{\theta}u'_r}{r} + u'_z \frac{\partial u'_{\theta}}{\partial z}\right] \hat{\theta} \dots (4.2)$$

$$+ \left[\frac{\partial u'_z}{\partial t} + u'_r \frac{\partial u'_z}{\partial r} + \frac{u'_{\theta}}{r} \frac{\partial u'_z}{\partial \theta} + u'_z \frac{\partial u'_z}{\partial z}\right] \hat{z}$$

We consider the vortex to be axisymmetric, so the variation of particle velocity with θ is neglected and due to steady state condition the time variation is also neglected. This reduces equation (4.2) to:

$$\frac{d\vec{u'}}{dt} = \left[u_r'\frac{\partial u_r'}{\partial r} - \frac{u_{\theta}'^2}{r} + u_z'\frac{\partial u_r'}{\partial z}\right]\hat{r} + \left[u_r'\frac{\partial u_{\theta}'}{\partial r} + \frac{u_{\theta}'u_r'}{r} + u_z'\frac{\partial u_{\theta}'}{\partial z}\right]\hat{\theta} + \left[u_r'\frac{\partial u_z'}{\partial r} + u_z'\frac{\partial u_z'}{\partial z}\right]\hat{z} \quad \dots (4.3)$$

Hence, as per Newton's second law, $M \frac{d\vec{u}}{dt} = F_p$, the equation of motions of sand grains in steady state can be written as

$$u_{r}^{\prime}\frac{\partial u_{r}^{\prime}}{\partial r}-\frac{u_{\theta}^{\prime}}{r}+u_{z}^{\prime}\frac{\partial u_{r}^{\prime}}{\partial z}=\frac{1}{8M}\pi\rho D^{2}(C_{D}+C_{L})\left[\left(u_{r}-u_{r}^{\prime}\right)^{2}+\left(u_{\theta}-u_{\theta}^{\prime}\right)^{2}+\left(u_{z}-u_{z}^{\prime}\right)^{2}\right]^{\frac{1}{2}}\left(u_{r}-u_{r}^{\prime}\right)\dots(4.4)$$

$$u_{r}^{\prime}\frac{\partial u_{\theta}^{\prime}}{\partial r} + \frac{u_{\theta}^{\prime}u_{r}^{\prime}}{r} + u_{z}^{\prime}\frac{\partial u_{\theta}^{\prime}}{\partial z} = \frac{1}{8M}\pi\rho D^{2}(C_{D} + C_{L})\left[\left(u_{r} - u_{r}^{\prime}\right)^{2} + \left(u_{\theta} - u_{\theta}^{\prime}\right)^{2} + \left(u_{z} - u_{z}^{\prime}\right)^{2}\right]^{1/2}\left(u_{\theta} - u_{\theta}^{\prime}\right) \dots (4.5)$$

$$u_{r}^{\prime} \frac{\partial u_{z}^{\prime}}{\partial r} + u_{z}^{\prime} \frac{\partial u_{z}^{\prime}}{\partial z} = \frac{1}{8M} \pi \rho D^{2} (C_{D} + C_{L}) \Big[(u_{r} - u_{r}^{\prime})^{2} + (u_{\theta} - u_{\theta}^{\prime})^{2} + (u_{z} - u_{z}^{\prime})^{2} \Big]^{\frac{1}{2}} (u_{z} - u_{z}^{\prime}) - \frac{1}{6M} \pi D^{3} (\rho_{p} - \rho) g \qquad \dots (4.6)$$

where, u'_r , u'_{θ} and u'_z are cylindrical velocity components of the particles (radial, tangential and vertical respectively), M is the mass of each particle and u_r , u_{θ} , u_z are the components of the vortex winds. The value of constants used in equations (4.4) to (4.6) are taken for the Martian atmosphere as reported in the literature [40, 277], i.e., $C_D = 0.4$, $g = 3.72 m s^{-2}$, $M = 3.1 \times 10^{-14} kg$, $\rho = 0.0176 kg.m^{-3}$ and $\rho_p =$ 2730 kg.m⁻³. The value of C_L varies with the particle size and the viscosity of the medium, so it is calculated accordingly in our model ([278], equation (6)).

Once we obtain the particle velocities, we compute the dust particle distribution throughout the length of the vortex, by numerically solving the following mass conservation equation,

$$\frac{1}{r}\frac{\partial(ru_{r}\phi)}{\partial r} + \frac{1}{r}\frac{\partial(u_{\theta}\phi)}{\partial \theta} + \frac{\partial(u_{z}\phi)}{\partial z} = 0 \qquad \dots (4.7)$$

$$\vec{\nabla} \cdot \left(\phi \vec{u'} \right) = 0 \qquad \dots (4.8)$$

where, ϕ is the dust particle concentration and $\vec{u'}$ is the effective total particle velocity, i.e., $\vec{u'} = \sqrt{(u_r'^2 + u_z'^2)}$. The particle's tangential velocity u'_{θ} is not considered for calculating the dust concentration. The radial component of velocity is providing the drag force and the vertical component of velocity is providing the lift force to the particles. But the tangential velocity is maintaining the particle at the given radius and height and hence does not contribute to lifting. Moreover, since we are considering our model domain to be an azimuthally symmetric cylinder, we can neglect the contribution of tangential component of the particle velocity.

Computing the airflow of the dust devil vortex is beyond the scope of this work, as we are considering the mature stage of the vortex in steady state. Thus to determine the velocity components for the vortex system (u_r, u_θ, u_z) , we need an analytical model for the vortex velocities. There are several analytical models for the wind velocity components within a vortex, of which the Burgers vortex and Rankine vortex are closest to observations. However, the drawback of the Rankine vortex is that it has no radial and vertical velocity components and also overestimates the velocity near the core of the vortex. Whereas, the limitation of Burgers vortex is that it assumes a linear variation of the radial velocity with the radius. In our work, we use an empirical model, for the vortex winds [279].

$$u_r = \frac{-6\nu (r/r_c)^3}{r_c \left(1 + (r/r_c)^4\right)} \qquad \dots (4.9)$$

$$u_{\theta} = \frac{\Gamma r / r_c^2}{\sqrt{1 + (r/r_c)^2}} \qquad \dots (4.10)$$

$$u_{z} = \frac{24\nu z (r/r_{c})^{2}}{r_{c}^{2} (1 + (r/r_{c})^{4})^{2}} \qquad \dots (4.11)$$

where, u_r , u_{θ} and u_z are the radial, tangential and vertical velocity components of the vortex, v is the atmospheric viscosity, Γ is the circulation strength of the vortex and r_c is the radius of the vortex. Several vortex forming experiments have been performed in a closed cylindrical chamber, using air and water [279]. Results from these, show that the radial distribution of the azimuthal velocity is not dependent on the method of their production, while a new empirical formula is derived for the tangential velocity which fit well with the observations. The radial and vertical velocity components are then derived from the equations of motion. This provides a solution for the velocity components which are bounded over the infinite domain, unlike any other theoretical model. Since the Vatistas model [279] uses the observation of several vortices to determine the wind field within it, it thereby incorporates the turbulence in the wind field solution implicitly.

We take the radius of the steady state vortex in our calculation as 10 m and height as 1000 m, since the observations state that dust devils are few kilometers in height and tens of meters in radius [115]. Recently, HiRISE has observed a dust devil in October 2019 at latitude 33° and longitude 202° E and Ls = 87° (Northern Spring), which was 50 m in width and a height of 650 m, on the dust-covered, volcanic plains of Amazonis Planitia [https://www.uahirise.org/ESP_061787_2140]. As mentioned in chapter 3, dust devils have also been imaged by the Mars Color Camera (MCC) onboard ISRO's Mars Orbiter Mission (MOM). Five dust devils were observed by the MCC on 07 November 2016 with 25.52 m resolution at spacecraft altitude of 490.66 km. The altitudes of these dust devils which were estimated using shadow method, varied from ~0.5 to 1.9 km [128]. Another study by Stanzel et al., (2008) [125] is based on the detection of dust devils by images recorded by HRSC on-board ESA Mars Express Orbiter. They provide an estimate of the diameter and height of the observed vortices, which varies from few meters to tens of meters in diameter and hundreds of meter to few thousand meter in height [125, 212], and our dust devil size also lies within the estimate.

We use the finite difference method to solve the non-linear coupled partial differential equations (4.4) – (4.8). We take evenly spaced grid for the r and z coordinates with a grid spacing of 0.1 m. The boundary conditions for numerical solution of equations (4.4) to (4.6) are such that the particle velocities at the center of the vortex become zero, since the velocity at the center of the vortex is negligible. At the lower boundary (surface), $u'_r = 0$, $u'_{\theta} = 0$, $u'_z = 0.01 m/s$. For determining the wind velocity components within a steady state vortex (equations (4.9) to (4.11)), we parameterize the viscosity v inside the spatial domain of the vortex. The value of the circulation strength, Γ can be considered constant throughout the vortex, whereas a variation in the value of v is expected in the vertical direction, due to high turbulence inside the vortex. As can be noticed from equation (4.11), the vertical velocity is solely dependent on v and not on Γ . We thus use simulated vertical winds of a dust devil from a mesoscale model [190] to determine the functional dependence of viscosity (v) on the height of the vortex. From this mesoscale model, we consider the variation of u_z/z with vertical height z, to obtain a higher order polynomial fit of the form:

$$\frac{v}{v_o} = \frac{\sqrt{z} \cdot \exp(-a(z-b)^2)}{\max[\sqrt{z} \cdot \exp(-a(z-b)^2)]} \dots (4.12)$$

where a = 0.0001 m^{-2} and b = 30 m are constants (fitting parameters representing the width and amplitude of the vertical viscosity variation), and v_o is the peaking viscosity. We choose the values of Γ and v_o in such a way that the total wind speed (sum of all three components) in a vortex achieves the minimum threshold value for wind lifting. In Martian atmosphere the threshold tangential wind speed required to lift dust of micron size is ~30 m/s [135, 149, 168], compared to ~20 m/s on Earth [275]. Using this threshold wind speed, we arrive at a value of the vortex strength, Γ = $450 m^2 s^{-1}$ and the peak viscosity, $v_o = 0.1 m^2 s^{-1}$. We perform simulations for particles with diameter 1.0 µm, 2.0 µm, 3.6 µm and 5.0 µm, so as to incorporate the major size range of particles that satisfy the gamma size distribution in the Martian atmosphere [280, 281]. The effective radius for this particle size distribution is 1.8 µm, which is the reason for us to take the above mentioned value of threshold wind speed.

We use equation (4.8) to determine the concentration of particles throughout the vortex, by calculating the effective total velocity (u') of the particles. We limit the solution of equation (4.8) by fixing the particle density at the upper boundary of the devil (1000 m). Sheel and Haider, (2016) [64] have calculated dust concentrations for different dust scenarios (background haze, local and global dust storms), assuming a gamma distribution with effective radius and variance of 1.8 µm and 0.3 respectively. Using their density for an effective radius of 1.8 µm, we calculate the number density of particles for other radii at an altitude of 1 km, balancing the atmospheric buoyancy of all particles.

4.3 **Results and discussions**

Observations of dust distribution inside dust devils are not available for Mars, whereas such measurements are available for the Earth's atmosphere. Raack et al., (2018) [264] sampled the vertical particle size distribution of two active dust devils having different sizes and intensities during a field campaign in the Sahara Desert (Morocco). Using these observations, they derived the relative lifted particle loads. Their measurements show that the majority of the particles in suspension were lifted only within the first meter of the dust devils (~76.5 wt % and ~89 wt % of the relative particle load respectively) with an exponential decrease of relative particle load with height. They also found a decreasing trend of particle sizes with altitude within the dust devils. The distribution of dust aerosols in dust devils has also been studied based on a field observation in Taklimakan desert, China [282]. They applied the Digital Optical Method (DOM) using digital cameras to quantify the dust opacity within the dust devils. Their results show that the opacity (which is directly proportional to number of grains) decreased monotonically with height, within the dust devils. So to validate our model, we first simulate the dust distribution by taking conditions for the Earth's atmosphere in our model, and our results show an exponential decrease of particle load with height, similar to the observations discussed. Thus, we can extend the use of our model for studying dust characteristics in dust devils on Mars.

As already discussed in section 4.2, we simulate the spatial distribution of dust concentration for four different sizes of dust particles, with radius of 0.5, 1.0, 1.8 and

 $2.5 \ \mu m$, as shown in Figure 4.1. We have chosen these sizes so as to cover the entire gamma distribution of dust particles.



Figure 4.1: Dust number density (cm⁻³ or #/cc) at different heights for four sizes of dust particles, with diameter d = 1.0, 2.0, 3.6 and $5.0 \mu m$.

It can be seen from Figure 4.1 that at each height, the maximum number density of the lifted dust is attained by particles with radius = $0.5 \ \mu m$. The larger the size of the particle, the smaller is the number density. The number density of larger size particles also decreases sharply as we move higher, indicating that larger grains tend to be lifted up to lower heights. Several observations of dust devils on both Earth and Mars indicate that dust particles with larger sizes are only lifted within the first few meters above ground [283, 284]. The size distribution of Martian dust follows a gamma distribution, while the predominant particle radius in the atmosphere is found to be 0.5 μm [64, 78,

280], which is also seen in Figure 4.1 of this work. Thus, in the remainder of this section, we will discuss our simulation results for particles with radius = $0.5 \,\mu m$.

Figure 4.2 shows the variation of the radial, tangential and vertical components of the particle velocity with height, for our vortex at two different radial distances from the center of the vortex (r = 2 m and at the boundary of the vortex).



Figure 4.2: Variation with height of (a) radial and tangential and (b) vertical components of the particle velocity, for two different radial distances from the center of the vortex.

The radial and tangential velocity components are primarily responsible for the sustenance of the dust in the vortex. We observe that these velocity components rise sharply as we move upwards from the surface. These velocities peak at ~ 60 m, thereafter saturating above ~ 100 m. This implies that the vortex is mostly turbulent in the lower part. The vertical velocity component leads to the lifting of particles along

the height of vortex, with the lifting velocity of the particle being counter-balanced by gravity. Figure 4.2(b) shows that as we move higher the vertical velocity initially increases signifying the upward movement of particles, but starts decreasing above a height of ~100 m and slowly approaches zero. This is in accordance with the vertical distribution of viscosity. We also observe that the particle velocities increase as we move towards the boundary of the vortex. The vortex has maximum strength at the boundary and hence the particle velocities are also the strongest. This distribution of the particle velocity components influences the particle distribution, which is discussed next.

Figure 4.3 shows the variation of particle number density (cm⁻³) with the height of the devil, for the two distances from the center of the vortex. The vertical height in the plot is restricted to 10 m since above this height, dust concentration saturates.



Figure 4.3: Variation of particle number density (#/cc) with height for two different radial distances from center of the vortex.

The dust density decreases exponentially with height, implying that though dust devils are understood to lift much dust from surface in comparison to near surface winds; the dust is not lifted to large heights. The dust devil raises most of the dust in the lower tens of meter, forming a "sand skirt" [48, 138]. The dust devil can sustain for a brief period of time (1–10 minutes on average) [210] and hence the suspension of particles also exists for small duration. The decrease in dust concentration with increasing height results from a competing effect of gravity and a change in vertical winds for particles. The larger particles tend to be lifted up to lower heights since their mass is high and gravity tends to settle it down faster. Moreover, it becomes easy for smaller particles to be entrained by the wind to higher heights and also remain suspended in air due to smaller mass.

The concentration of particles during a dust devil condition has been observed to be ~1500 cm⁻³ [281]. It matches well with our model results, which show a concentration of ~1400 cm⁻³ near the surface. We use the simulated particle densities and the vortex wind speed in our model, to predict the flux of dust lifted near the surface. An average mass dust flux over the entire range of the vortex near surface is estimated to be ~5 × 10⁻⁵ kgm⁻²s⁻¹. This is well in agreement with the laboratory simulation result of the dust flux within a dust devil whose value varies between ~1 × 10^{-5} to ~3 × 10^{-2} kgm⁻²s⁻¹ [83]. Our result also matches with the dust flux estimated in the range ~6×10⁻⁴ kgm⁻²s⁻¹ to ~5×10⁻³ kgm⁻²s⁻¹ at the Mars Pathfinder site using optical images of the dust devils taken from the lander [130]. Reiss et al., (2014) [211], measured the optical depth of three separate dust devils using HiRISE images and estimated a dust flux in the range ~3.8 × 10⁻⁷ to ~1.2 × 10⁻³ kgm⁻²s⁻¹. Figure 4.4 shows the variation of dust concentration near the surface with the radial distance from the center of the dust devil. A convective vortex is characterized by its corresponding pressure drop between the center of the vortex and the background, generated by the swirling motion of winds.



Figure 4.4: Variation of surface dust density (#/cc), with radial distance from the center of the vortex.

A higher magnitude of pressure drop indicates a stronger vortex with higher tangential velocity and is thus more likely to form a dust devil [143, 270]. The pressure is least at the center of the vortex and increases towards the boundary of the vortex, thus making pressure drop (pressure difference between the center and the boundary of the vortex) highest at the boundary [190, 274]. Hence, moving towards the edge of the vortex, the winds become stronger and reach their maxima at the boundary, thus lifting more particles at 10 m, the boundary of the vortex [51, 195]. Outside the boundary, the effect

of the vortex diminishes and hence the vortex winds also decrease [279]. Hence the particle concentration decreases rapidly due to fall in particle velocity.

4.3.1 Effects of model parameters

We study the effect of variation of parametric variables (Γ and v_o) on our model simulated particle velocities and concentration. As already mentioned in section 4.2, these two parametric variables are so chosen that the vortex tangential winds reach the threshold minimum of ~30 m/s for dust devil formation. In this section, we consider different values of these parameters, so as to achieve a threshold wind velocities of ~10 m/s and ~70 m/s also, as reported in the literature [115]. We evaluate the vortex wind velocities (equations (4.9) – (4.11)) and the particle velocities (equations (4.4) – (4.6)) using these parametric values.

4.3.1.1 Variations in circulation strength (Γ)

We consider two other values of the vortex strength, $\Gamma = 100 m^2 s^{-1}$ and $1000 m^2 s^{-1}$ respectively and fix $v_o = 0.1 m^2 s^{-1}$. Since the particles' velocities are guided by the wind velocities in the vortex, the change in wind velocity will affect the particles' velocity. Figure 4.5 shows the variation of radial and tangential components of the particle velocities, at the boundary of the vortex.

The trend of the velocity does not change but the magnitude changes. Since the particle velocities are guided by the vortex wind velocities, the effect of Γ on the particle velocity can be seen in light of the variations in wind velocity. An alteration in Γ leads to a significant change in the tangential component of the wind velocity (due to the

direct dependence on Γ) and hence a corresponding change in tangential particle velocity in Figure 4.5.



Figure 4.5: Variation with height of radial and tangential components of the particle velocity, for different values of circulation strength Γ at the boundary of the vortex.

Despite a lack of direct dependence of radial wind velocity on the circulation strength, this component of the velocity has an effect on the vortex sustenance. For a vortex to sustain, the radial wind is balanced by the tangential wind of the vortex. Hence the effect of Γ on the tangential particle velocity translates to a significant effect on the radial component as well. The vertical wind velocity is independent of the circulation strength (equation (4.11)) and also since it depends on the pressure drop at the vortex center, it is not influenced by the tangential wind velocity. Therefore, we do not observe
any variation in the vertical component of the particle velocity with change in Γ (and hence not shown in Figure 4.5).

Figure 4.6 shows the variation of particle number density with the height of the devil, at the boundary of the vortex for different circulation strengths Γ . The increase in dust density with increasing vortex strength is a reflection of the corresponding change in the effective total particle velocity, with major contributions from the radial and tangential components.



Figure 4.6: Variation of dust number density (#/cc) with height for different values of circulation strength Γ at the boundary of vortex.

4.3.1.2 Variations in viscosity (ν_o)

We take two other values of v_o to be 0.01 $m^2 s^{-1}$ and 1.0 $m^2 s^{-1}$ respectively and fix the value of $\Gamma = 450 \ m^2 s^{-1}$. Figures 4.7(a) and 4.7(b) show the variation of radial, tangential and vertical components of the particle velocities at the boundary of the vortex for different values of v_o . We do not find any significant change in the tangential component of the particle velocity for changing viscosity (at a constant value of Γ). Consequently, we do not find any significant change in the radial component of the particle velocity as well. But the vertical component of the particle velocity is significantly modified for higher values of v_o , which is due to a similar effect in the vertical wind velocity.



Figure 4.7: Variation with height of (a) radial and tangential and (b) vertical components of the particle velocity, for different values of viscosity v_o at the boundary of the vortex.

Figure 4.8 shows the variation of particle concentration for different values of viscosity at the boundary of the vortex. No significant change is observed in the particle load due to a change in the viscosity. It can be understood from the fact that a change

in viscosity leads to a significant change in the vertical component of the particle velocity, but not in the other two components. Hence the effective total particle velocity does not vary much for the three different values of viscosity.



Figure 4.8: Variation of particle number density (#/cc) with height for different values of viscosity v_0 at the boundary of vortex.

4.3.2 Optical and thermal effects of lifted dust

Dust devils are believed to be responsible for maintenance of the haze in the atmosphere [48, 131, 138, 142]. The global haze has a strong influence on the Martian climate since it controls the amount of sunlight reaching the surface, and also leads to heating of the atmosphere [58, 85, 86, 166]. The optical depth and heating rates due to dust have been well studied for different seasons and regions of the Martian atmosphere [64]. Here we estimate these for the simulated dust devil in the boundary layer. The

optical depth τ , which is a proxy for the dust loading in the devil, is a measure of the attenuation of solar radiation by the dust in the devil and can be defined as

$$\tau = \int_{z_0}^{z_1} \sigma n(z) dz \qquad \dots (4.13)$$

where, σ is the extinction cross-section of dust aerosol and n(z) is the number density of the dust at height z, the integral being from the surface (*z*₀) to the top of the dust devil (*z*₁). The value of extinction cross-section of Martian dust is approximately 1.84 × $10^{-11} m^2$ for visible wavelengths [60, 285]. Using these values, we obtain a total optical depth of ~0.2 for our simulated dust devil. Optical depths in the visible region have been estimated from three separate dust devils using HiRISE images in the range 0.3 - 1.2 [211]. As the dust density decreases with altitude, so does the corresponding optical depth τ , and at a given altitude the aerosol heating rate (K s⁻¹) can be given as [10],

Heating Rate =
$$\frac{g}{C_p} f_a F_{\Theta p} \frac{2\tau_o}{P_o} \exp(-2\tau)$$
 ...(4.14)

where, g is the acceleration due to gravity, C_p is specific heat at constant pressure, P_o is the surface pressure, f_a is the absorption fraction of the solar irradiance at the top of the atmosphere ($F_{\odot p}$), and τ_0 is the total optical depth of the dust devil at the surface. We assume that ~10 % of incoming solar flux is absorbed by the dust [265] and hence f_a is 0.1. Using our estimated optical depth $\tau_0 = 0.2$, the near surface heating rate is ~0.01 Ks⁻¹. This is close to the value of temperature rise reported from a study of solar heating in a dust devil from the images observed by Mars Orbiter Camera [286]. If the dust devil remains active for even one minute with the same intensity, then it can lead to an increase in temperature by approximately 0.8 K. This local heating may affect the dynamics of the atmosphere from a synoptic scale to a larger scale [166], since dust influences the radiative energy budget directly by scattering and absorbing solar radiation [166, 287]. The heating of the atmosphere by dust may generate more dust devils due to positive feedback effect.

4.4 Conclusion

This study focuses on the numerical modeling of dust lifting within a steady state dust devil in the Martian atmosphere. This is a unique study on determination of spatial distribution of dust number density within a dust devil. The numerical model takes into consideration the vortex wind equations within the steady state dust devil [279] parameterized in terms of the circulation strength Γ and viscosity ν . We consider the drag force, lift force and gravitational force in the equations of motion for dust to determine the velocity of the particles. The continuity equation is then used to determine the concentration distribution of dust particles within the devil. Our results indicate that major particle load (~1400 cm⁻³) in the steady state dust devil is near the surface and at the boundary of the vortex leading to a sand skirt, which is consistent with observations. The dust number density decreases as we move in the vertical direction, with lighter particles lifting to higher heights. The height up to which dust is lifted significantly, is within the first 10 m from the surface, above which the concentration tends to saturate. This means that though dust devils are understood to lift much dust from surface in comparison to near surface winds, the dust is not lifted

to large heights. The near surface dust flux estimated by our model (~ 5×10^{-5} kgm⁻²s⁻¹), is also consistent with observations [130] and laboratory simulations [83]. Amongst the parameters of our model, the dust loading is more sensitive to variations in the circulation strength (Γ) than to variations in viscosity (ν). We report an optical depth of 0.2 and a heating rate of 0.01 Ks^{-1} due to our simulated dust devil. Thus the heating due to suspended dust particles in the Martian dust devils can affect the boundary layer processes and lead to more occurrences of the dust devils. The result which we obtain for the dust particle distribution within a steady state dust devil is dependent on the size of vortex in consideration. The estimates will vary once we change the size of the vortex, which is also evident from the observational and laboratory estimates of dust flux and optical depth for dust devils on Mars.

Dust lifting is a sub-grid scale process at the resolution of a general circulation model (GCM) of the Martian atmosphere, therefore it is parameterized. One such parameterization is based on the vortex tangential wind speed v_t being more than the threshold, in order to lift dust. In this scheme, the dust flux is parameterized in terms of the dust devil lifting efficiency α_D as dust flux = $\alpha_D(\rho v_t^2 - 15)/g$ [46]. Here ρ is the density of air and g is the acceleration due to gravity. The uncertainty surrounding the actual quantity of dust that devils are able to lift leads to α_D being a tunable parameter in Martian GCMs. Using the dust flux and the vortex average tangential velocity estimated from our simulations, we obtain the value of $\alpha_D = 1.4 \times 10^{-4} \text{ s}^{-1}$. Thus, our simulations are able to provide constraints on the parameter space of α_D . Our study provides dust densities and particle velocities within a dust devil in the boundary layer, which could be important inputs to understand the effect of dust on the radiative and dynamical processes in the boundary layer. Our results may be useful in future studies to estimate the contribution of dust devils to the global dust entrainment into the Martian atmosphere.

Chapter 5 Electric Field Development within a Dust Devil

As discussed in chapter 4, the winds in the dust devils are strong enough to lift dust from the surface and distribute it throughout the spatial extent of the dust devil. In this process, the dust particles collide with each other and are known to generate and transfer electric charges by triboelectric mechanism [288–295]. The dust particles get charged based on their size and composition [291, 296, 297]. Mass stratification of dust occurs within a dust devil on the basis of particle size, due to which an electrostatic field is developed. In this chapter, we will discuss the electric field generation within a dust devil in the Martian atmosphere. We shall first describe the observations of the electric field in terrestrial dust devils due to the lack of such observations on Mars. After that, we provide a physics insight of the electric field generation within the dust devil in the Martian atmosphere based on analytical and numerical solutions.

5.1 Observations of electric field within a dust devil

Multiple observations of dust devils are made for Mars using both orbiters and surface based instruments, as described in chapter 3. Various field test campaigns have been conducted in different desert areas on Earth to measure the meteorological parameters (like pressure, temperature, wind, humidity) for dust devils and dust storms [107, 147, 288, 289, 298–308]. These terrestrial observations show that the electric fields within dust devils can vary from a few kV/m to ~100 kV/m [147, 303, 307–309]. Figure 5.1 shows the electric field within a dust devil observed during a field campaign in the West Sahara Desert [307]. A strong positive correlation between the lifted dust and the electric field intensity was found, indicating that all dusty events (dust devils) involve a similar dust electrification process [307].



Figure 5.1: Observation of the electric field within a dust devil during a field campaign in the West Saharan desert [Image credit: Esposito et al, 2016 [307]].

As mentioned above, the observations of the electric field within dust devils are available only for Earth, as of now. No rover or lander on Mars has recorded any direct measurement for the electric field within the dust devils.

5.2 Electrostatic system in the atmosphere

The electricity in the atmosphere depends on the concentration of charged particles and the meteorological processes involved in separating these charges to form a dipole like system [310, 311]. The outcome of electrification of the atmosphere is the motion of charged particles and ions under the influence of electric fields. The electrical discharges are also observed due to a substantial accumulation of these charges. The laboratory experiments of the electrical breakdown obtained by mixing sand grains in a low-pressure CO₂ environment show pre-glow and spark discharge [312–314]. The adhesion of the dust to the wheels of the Mars Pathfinder and Sojourner rovers are suggested to be electrostatic in origin and is indirect evidence of electrification in the Martian environment [315, 316]. Thus we have reasons to believe that the dust devils on Mars are also electrified. The galactic cosmic rays, X-Rays, and solar EUV rays are responsible for the generation of the ions and free electrons in the atmosphere of Mars at different altitudes, giving rise to an electrically conductive atmosphere (Ionosphere) [66, 317, 318]. Due to the thin atmosphere of Mars, the cosmic rays penetrate deep into the atmosphere and ionize neutral atoms and molecules up to the surface. The total electrical conductivity of the atmosphere is given as [319],

$$\sigma = e(\mu_{+}n_{+} + \mu_{-}n_{-} + \mu_{e}n_{e}) \qquad \dots (5.1)$$

where, n_+ , n_- and n_e are the densities of positive ions, negative ions and free electrons respectively, μ_+ , μ_- and μ_e are their associated electrical mobility.

Haider et al., (2010) [66] have developed an ion-dust aerosol model to compute the concentrations of ions and electrons and consequently study the electrical

conductivity on Mars. It was found that during the dust storms, the concentration of major ions considerably reduces due to attachment with dust. This leads to a decrease in the conductivity of the atmosphere in the presence of dust [66]. The conductivity of the Martian atmosphere varies with height, with a value of $\sim 8 \times 10^{-14} Sm^{-1}$ near the surface in presence of dust whereas $\sim 5 \times 10^{-12} Sm^{-1}$ near the surface in absence of dust [66]. The conductivity of Earth's atmosphere near the surface is $\sim 6.6 \times$ $10^{-14} Sm^{-1}$ [320], which is approximately two orders less than that of the Martian atmosphere in the absence of dust. The conductivity of Earth's atmosphere near the surface is less as compared to Mars' due to lack of ions and electrons in the lower atmosphere. Due to higher value of conductivity in Martian atmosphere, the charges developed inside the dust devils will reach relaxation early as compared to those on Earth. The charge relaxation timescale $(\tau = \frac{\varepsilon_0}{\sigma})$ is smaller for charges within a dust devil on Mars as compared to Earth, suggesting a faster charge decay. This will eventually lead to the generation of weaker electric field within dust devils in Martian atmosphere. This lower magnitude of the electric field is such that it does not reach the atmospheric breakdown magnitude. It makes the chances of observing lightning in dust devils on Mars less probable in comparison to the Earth. But once the Martian atmosphere is covered with dust, the conductivity of the atmosphere becomes comparable to its Earth counterpart. Hence, just like on Earth, we may find the development of strong electric field within dust devils in the Martian atmosphere.

5.2.1 Charging of dust particles

The basic principle of the particle charging on Mars is assumed to be similar to that in the Earth's atmosphere. The charge build-up on the dust particles occurs when they collide with each other, also known as "triboelectric charging." This is a contact electrification process occurring when two particles come in contact for a finite duration with each other and separate, leaving the constituent surfaces often charged – the charge transfer is usually equal in magnitude and opposite in nature [321-324]. Although triboelectric charging appears simple, the role of rubbing in charging the particles is unclear [325]. Three processes are mainly involved in tribocharging – electron transfer, ion transfer, and material transfer [323, 324]. The charge exchange via tribocharging also depends on the characteristic material properties, such as electrical conductivity, permittivity, the microstructure of the surface, and chemical structure [326]. A laboratory study of the charge and size distribution of particles by Kunkel, (1950) [327] found that the average charge is roughly proportional to the particle radius. Ette, (1971) [291] also had similar results for silica dust clouds, showing the larger and heavier particles to be positively charged and the smaller and lighter particles to be negatively charged. In the absence of tribocharging of the dust particles in the lower Martian atmosphere, the charges acquired by the particles (by conduction and induction methods) will not exceed 10 - 20 electron charges [328].

Thus, to simulate this effect of larger particles getting positively charged and smaller particles getting negatively charged, Melnik and Parrot, (1998) [296] assumed that, upon collision of two particles, the smaller particle obtained a femto-coulomb

(fC ~10⁻¹⁵ C) of negative charge for each micrometer of its radius while the larger particles take away an equal but opposite charge. Hence, the charge exchange per collision (Δq) between two particles may be given as [296],

$$\Delta q = (1fC/\mu m)r_s \qquad \dots (5.2)$$

where, r_s is the radius of the small particle. Another approach for estimating the charge exchange per collision between the two particles is given by its composition, i.e., the difference in the contact potentials between two materials [297]. They give the charge transfer to the larger particle as,

$$\Delta q = f_1 \Delta \varphi - (1 - f_2) q_{tot} \qquad \dots (5.3)$$

where $\Delta \varphi$ is the difference between surface triboelectric potentials of the particles, q_{tot} is the total charge on both particles, and f_1 and f_2 are the constants depending on the mutual capacitances of the two particles (i.e., their geometry). Equation (5.3) shows that even if the two particles have the same composition ($\Delta \varphi = 0$), there will be a charge exchange occurring solely due to the existence of the particle's mutual capacitance. In the case when $\Delta \varphi \neq 0$ and $r_L \gg r_S$, equation (5.3) can be rewritten as [297],

$$\Delta q \approx 2668 \left(\Delta \varphi / 2V \right) (r_f / 0.5 \mu m) e \qquad \dots (5.4)$$

where, r_f is the reduced radius ($r_f = (r_L^{-1} + r_S^{-1})^{-1} \sim r_S$), and *e* is the electronic charge. These equations ((5.2) and (5.4)) will be used further in the calculation of the electric field within a dust devil for the Martian atmosphere. We take $\Delta \varphi$ to be 2V since it is consistent with the easily lifted iron/silica particle mix on Mars [297, 329].

5.3 Analytical solutions to the electric field within a dust devil

Using the particle charging processes described above, we do an analytical study of electric field generation within dust devils on Mars. This methodology is similar to the electrostatic model developed for induction-produced electric fields in terrestrial thunderstorms [330–332]. We assume a dust devil with vertical charge separation resulting from the vertical transport of charged dust particles. We also assume the dust devil to maintain its size throughout the phase of electrostatic field generation. Hence, the electric field E, developing in a dust devil is given in terms of continuity equation as,

$$\frac{dE}{dt} = -\frac{J}{\varepsilon_o} \qquad \dots (5.5)$$

where, E is the electric field and ϵ_o is the free space permittivity. J corresponds to the current density that can be given as,

$$J = n_L Q_L v_L + n_S Q_S v_S + \sigma E \qquad \dots (5.6)$$

where, $n_{L,S}$ is the number of large and small particle concentration, $Q_{L,S}$ is the charge on large and small particles, $v_{L,S}$ is the vertical velocity of large and small particles, and σ is the conductivity of the atmosphere. The term σE is the dissipation current, which limits the growth of the electric field and thus leads to saturation. Though there will be a buildup of charges within the dust devils, we expect the overall charge in the devil to have a net value of zero to maintain charge neutrality, thus $n_L Q_L = -n_S Q_S$. The current density now reduces to,

$$J = n_I Q_I \Delta V + \sigma E \qquad \dots (5.7)$$

where, $\Delta V = v_L - v_S < 0$. Substituting equation (5.7) into equation (5.5) and timedifferentiating it yields,

$$E^{\prime\prime} + \frac{\sigma}{\varepsilon_o} E^{\prime} = -\frac{n_L \Delta V}{\varepsilon_o} Q_L^{\prime} \qquad \dots (5.8)$$

where prime over variables indicates time differentiation (d/dt). Q'_L is the time rate of charge increase on large grain and can be expressed as [309]

$$Q_L' = \upsilon \Delta q, \qquad \dots (5.9)$$

where Δq is charge exchange per collision and υ is collision frequency given by $\upsilon = \pi r_L^2 \Delta V n_s$. We use equation (5.4) to derive the analytical solution to equation (5.8) considering the electric field and its first-order time differentiation to be zero at t = 0. We get

$$E = -\frac{n_L \upsilon \Delta V \Delta q}{\varepsilon_o} \left[t \frac{\varepsilon_o}{\sigma} - \left(\frac{\varepsilon_o}{\sigma}\right)^2 \left(1 - e^{-\sigma t/\varepsilon_o}\right) \right] \qquad \dots (5.10)$$

The size of dust particles which can be raised by dust devils in the Martian atmosphere varies between 1 μ m to 50 μ m [333–335]. Hence, we take the radius of smaller dust particles to be 1 μ m with a number concentration of 50 cm⁻³, and the larger dust particle to be 20 μ m with a number concentration of 1 cm⁻³ for the calculation of the electric field [336]. Greeley et al., (2010) [134] discuss the observations of dust devils as recorded by Spirit rover in the Gusev crater. They report the median vertical velocity of the dust devils to lie in the range of 1 – 1.6 ms⁻¹, with maximum value to be

~ 17 ms^{-1} . Hence, we take the value of vertical wind to be ~ 5 ms^{-1} (mean value) and then use equation (17) of Farrell et al., (2006) [336] to determine the value of $\Delta V \approx$ $-3 ms^{-1}$. Figure 5.2 represents the variation of the electric field within a dust devil for two cases, viz., the dusty and non-dusty Martian atmosphere (by taking conductivity in respective scenarios, discussed in section 5.2). This is obtained from equation (5.10) using the parameters described earlier.



Figure 5.2: Variation of electric field with time in dusty and non-dusty scenario of Martian atmosphere.

Figure 5.2 suggests that the electric field build-up is stronger in the presence of dust in the atmosphere. The reason for this lies in the fact that during the presence of dust, the conductivity of the atmosphere decreases by two orders as compared to the non-dusty atmosphere [318]. Even when a moderate electric field is developed in the Martian atmosphere, a substantial dissipation current is generated, which increases exponentially with increasing electric field. Thus, a significant competing dissipation

current develops in the atmosphere that acts to deplete the development of large separated charge centers and reduces the electric field. The electric breakdown value for the Martian lower atmosphere is ~20 kVm^{-1} [296, 329, 337]. We can see that in the initial 30 seconds, the electric field reaches the breakdown value in a dusty atmosphere, whereas it takes more than 100 seconds to reach the same in a non-dusty atmosphere. Thus, the chance to detect lightning inside a dust devil on Mars is higher if the atmosphere is covered with dust. This result will be valid only if we consider the dust devil to maintain a steady state with the above mentioned parameters throughout the time. If we reduce the value of ΔV to $-1 ms^{-1}$ (i.e., if we change the vertical wind velocity or the sizes of particles in consideration), we observe the magnitude of the electric field to become ~1 kVm^{-1} for non-dusty scenario and ~20 kVm^{-1} for dusty scenario after 100 seconds. This also suggests that it will take a longer time to reach breakdown in order to observe lightning in the dust devil. Hence, the role of velocity difference between particles also plays a significant role in the development and growth of the electric field within a dust devil.

In the above scenario the effect of charge relaxation due to conduction in the atmosphere (σQ_L) has not been accounted. By charge relaxation, we mean the tendency of the system to limit the growth of charge on the particles and help the system reach equilibrium. Earlier, we had considered that there is a constant rate of charging of dust present inside the dust devil. We ignored the fact that the timescales for particle charge relaxation may be of the same order, or sometimes exceed the currents generated from collisions between the particles. This is incorporated in the equation by considering the grain charge relaxation, as [337]

$$Q_L' + \frac{\sigma}{\varepsilon_o} Q_L = \upsilon \Delta q \qquad \dots (5.11)$$

The analytical solution to equation (5.11) is obtained by taking initial condition $Q_L(t=0) = 0$,

$$Q_L = \upsilon \Delta q \frac{\varepsilon_o}{\sigma} \left(1 - e^{-\sigma t/\varepsilon_o} \right) \qquad \dots (5.12)$$

Now, substituting equation (5.12) into equation (5.7) and equation (5.5), we get,

$$E' + \frac{\sigma}{\varepsilon_o} E = -\frac{n_L \upsilon \Delta V \Delta q}{\sigma} \left(1 - e^{-\sigma t/\varepsilon_o} \right) \qquad \dots (5.13)$$

We use equation (5.4) to derive the analytical solution to equation (5.13) considering the electric field and its first-order time differentiation to be zero at t = 0. We get

$$E = \frac{n_L \upsilon \Delta V \Delta q}{\sigma} \left[t e^{-\sigma t/\varepsilon_o} - \frac{\varepsilon_o}{\sigma} \left(1 - e^{-\sigma t/\varepsilon_o} \right) \right] \qquad \dots (5.14)$$

Using the same values for each parameter as defined earlier, we obtain the plot showing the variation of the electric field in the presence of charge relaxation for the dusty and non-dusty scenario.



Figure 5.3: Variation of electric field with time in presence and absence of charge relaxation for (a) Non-dusty scenario, and (b) Dusty scenario of Martian atmosphere.

In figure 5.3 we observe that due to charge relaxation the growth of electric field is affected, with a reduction in its maximum value, for both dusty and non-dusty scenarios. In non-dusty scenario, the conductivity is high, which leads to a faster relaxation and transfer of charges among the particles, thus leading to an early saturation of the electric field. Whereas in the dusty scenario, the atmospheric conductivity is lower, thus leading to a higher timescale required for charge relaxation as compared to non-dusty scenario. Due to this the charge relaxation becomes slow with time and makes electric field reach saturation late as compared to its non-dusty counterpart. The inclusion of relaxation makes the electric field to saturate at a lower value in a longer time, as compared to its earlier counterpart in which charge relaxation was ignored.

5.4 Numerical solution to the electric field within a dust devil

The analytical models discussed in the previous section provides a useful understanding of the electric field generation within the dust devils in the Martian atmosphere. However, for a better physics insight of this phenomenon, we perform a numerical simulation of the electric field generation within a finite-sized dust devil in the Martian atmosphere. Figure 5.4 shows a schematic diagram of our computational domain.



Figure 5.4: A schematic diagram of our computational domain.

We assume our system to be in a cylindrical coordinate system with symmetry about the z-axis. The dust devil is at the center of the domain, which is modeled by a cylinder of radius $r_o = 5 m$. the outer region is modeled by a cylinder of radius $20r_o =$ 100 m, and the height of the dust devil is taken to be 100 m. We solve the following basic electrostatic equations to determine the electric field.

$$\vec{\nabla}.\vec{E} = \frac{\rho}{\varepsilon_o} \qquad \dots (5.15)$$

$$\frac{\partial \rho}{\partial t} + \frac{\sigma \rho}{\varepsilon_o} = 0 \qquad \dots (5.16)$$

Where ρ is the charge density within the dust devil. We use the finite difference method to numerically solve these two equations within a dust devil. We perform a two-step simulation - firstly, to obtain the dust devil induced electric field and secondly, to obtain the total charge relaxation at each time step. For doing so, we consider the potential at the domain boundary to be zero. We also use an incremental charge density per unit time step of $\sim 3.14 \times 10^{-9} Cm^{-3}s^{-1}$ [296] as the input of the dipole charge configuration. As described earlier, the larger particles tend to charge positively and stay close to the ground, whereas the smaller particles tend to charge negatively and move to the top of the devil. In earlier studies [296, 337], an equal distribution of charge within the dust devil was assumed to solve for the electric field within the domain. They assumed the bottom half of the devil to be positively charged and the upper half to be negatively charged, i.e., a dipole configuration. However, in real scenario, this is not the case, and the number density of the particles decreases rapidly with altitude. The number density of the larger particles falls sharply than the number density of the smaller particles. This makes the larger particles to be dominant in the lower part of the dust devil, and smaller particles reach higher altitudes. But throughout the vertical axis, the total density of particles within the dust devil decreases with altitude. The charge density at each altitude is assumed to fall as r^4 along the radial distance, such that the charge density only remains present within the dust devil in our calculation domain. At each time, we compute the electric potential first for the obtained charge density within the domain using Poisson's equation. Then, using this electric potential, we calculate the corresponding electric field. Figure 5.5 shows a plot of the variation of charge density at the center of our calculation domain with time.



Figure 5.5: Variation of charge density at the center of our calculation domain with time.

Due to the higher value of conductivity on Mars, the charge density within the dust devil saturates faster and achieves a lower saturation value as compared to its Earth counterpart. This also means that the strength of the electric field developed in the Martian atmosphere for a particular size of dust devil will always be less in comparison to its terrestrial counterpart. Figure 5.6 shows a 2D plot of the charge density within our calculation domain at 500 s (~8 min).



Figure 5.6: Charge density distribution within our calculation domain at 500 s. The color bar on the right represents the magnitude of charge density in Cm^{-3} .

It can be seen from figure 5.6 that the positive charges are concentrated in the lower part of the dust devil and cover a smaller volume as compared to the negative charges. The charge density is also decreasing with altitude and reach a magnitude of zero, specifying the effect of vertical exponential distribution of dust particles within a dust devil. Due to this configuration of charge density, the electric field is developed in the domain, as shown in figure 5.7.



Figure 5.7: Electric field distribution within our calculation domain at 500 s. The color bar on the right represents the magnitude of the electric field in Vm^{-1} .

The maximum magnitude of the electric field within the dust devil in our simulation comes as $\sim 5.5 \ kVm^{-1}$ at 500 s. The magnitude of the electric field within the dust devil develops in the same way as the charge density developed with time. This magnitude is much below the breakdown limit of the electric field in the Martian atmosphere. Whereas, when we simulate the electric field in our domain with dipole charge density configuration, i.e., bottom half is positively charged and upper half is negatively charged, we obtain the maximum electric field to be $\sim 20 \ kVm^{-1}$. Figure 5.8 shows the electric field distribution within our calculation domain at 500 s.



Figure 5.8: Electric field distribution within our calculation domain at 500 s. The color bar on the right represents the magnitude of the electric field in Vm^{-1} .

The dipolar charge distribution within dust devil cannot occur in real atmospheric scenario. There will always be a vertically decreasing distribution of particles in which equal vertical separation of charges cannot occur. Hence, it is tough to reach breakdown within a dust devil. Moreover, due to a vertical decreasing particle density within the dust devil, the electric field generated inside it also shows a similar trend. Its magnitude decreases as we move higher in altitude. Whereas for a dipolar configuration of particles we observe that the electric field is maximum at the center and top of the dust devil, i.e., where charge polarity is changing.

5.5 Conclusion

In this chapter, we have discussed the electric field generation within a dust devil in the Martian atmosphere. We started with a brief description of the measurements of the electric field as observed by various instruments within dusty events on Earth. It gives us an idea of the strength of the electric field being developed in the atmosphere due to dust devils. However, similar observations are not available for the Martian atmosphere. We then discussed the conductivity of the Martian atmosphere and noted that the atmosphere is more conducting in the absence of dust. The conductivity of the Martian atmosphere in the absence of dust is approximately two orders higher than Earth's conductivity, thus leading to faster charge relaxation on Mars as compared to Earth before reaching the breakdown. Thus, the chances of observing lightning in dust devils on Mars is less probable as compared to Earth. The charging of dust within the dust devils is accounted for by a process termed "tribocharging". The mechanism of tribocharging describing the time rate of dust particle charging is discussed in section 5.2.1. The charging rate obtained by these methods is used to study the development of the electric field in the dust devil using analytical approaches.

The analytical solution to the charge continuity and electrostatic equation suggests that the electric field build-up is stronger in the presence of dust in the atmosphere. The magnitude of the electric field reaches the breakdown value $(\sim 20 \ kV m^{-1})$ in initial 30s for a dusty scenario, whereas its value in non-dusty scenario stands as $\sim 3 \ kV m^{-1}$ in the same time. The reason for this is a decrease in conductivity

in the presence of dust in the atmosphere, leading to low dissipation current and high charge buildup. Later, we incorporate charge relaxation due to conduction in the atmosphere (σQ_L) in our analytical solution. After incorporating charge relaxation, we observe the electric field to saturate at lower value with time for both dusty (~100 kVm⁻¹) and non-dusty (~200 Vm⁻¹) scenarios. The only difference between these scenarios lies in the time taken for the electric field to saturate. In a non-dusty scenario, the electric field saturates fast as compared to its dusty scenario's counterpart leading to a faster relaxation and transfer of charges among the particles.

Even though an analytical solution provides a useful understanding of the electric field generation within the dust devils, more realistic insight into this physical phenomenon is given by its numerical simulation. An analytical solution only tells about the maximum electric field within a dust devils and cannot predict the distribution of electric field strength throughout the domain. Whereas, the numerical simulation of the electric field generation accounts for the 2D distribution of electric field strength. We calculate the electric field within our described domain $(10 \ m \times 100 \ m)$ by simultaneously solving the Poisson's equation and charge density continuity equation. We consider larger particles to stay more close to the ground and smaller particles to move higher. A vertically decreasing particle distribution is considered to determine the final charge distribution within the dust devil. We observe that in such a charge distribution, the maximum magnitude of the electric field is $\sim 5.5 \ kVm^{-1}$, whereas when we determine the electric field for dipole-like configuration we obtain the maximum electric field to be $\sim 20 \ kVm^{-1}$. A dipolar vertical charge distribution can not be realized within an atmospheric dust devil. Thus, the chances of detecting

lightning within dust devil in the Martian atmosphere becomes sparse. But once the size of dust devil is increased, we can expect an increase in the strength of electric field generated within the dust devil for a vertically decreasing number density of particles. So, in the Martian atmosphere larger size dust devils can generate electric field within it which can reach breakdown and cause lightning.

Chapter 6 Summary and Future Work

Dust is a fundamental component of the atmosphere on Mars, and has a strong impact on the atmosphere's thermal and dynamical state [46, 166]. Convective vortices and dust devils occur very frequently in the Martian climate system [124, 133] and are an efficient mechanism by which dust is entrained into the atmosphere. Thus, it becomes important to understand the processes involved in vortex formation and consequently dust lifting, which are still under research. The objective of this thesis is to understand the characteristics of convective vortices on Mars. My study is based on steady state vortex systems, and does not deal with its formation or decay phase. In this chapter, we present a summary of work done for this thesis, and their corresponding results. Later, we also suggest the future direction in which we would like to continue our investigation of dust devils in Martian atmosphere.

6.1 Summary of results

In chapter 2, we present an analytical solution to the Navier Stokes (NS) equation. We have derived a simple equation for the mean tangential velocity in a cylindrical co-ordinate system using NS equation and continuity equation. The derived equation represents the dependency of tangential velocity on distance from the center

of the cylinder, and the altitude. This equation (equation (2.40) of chapter 2) would be useful to estimate the variation of velocity with radial distance from the vortex center. Unlike other analytical solutions for convective vortices, our solution is dependent on the measureable atmospheric parameters like threshold friction velocity and roughness length of the atmosphere. A comparison with observed data substantiates the validity and applicability of equation (2.40) for vortex systems in planetary surface layers.

In chapter 3, we have analyzed the meteorological observations made by Curiosity rover for MY 33 and detected a total of 611 daytime convective vortices having pressure drop $\Delta p > 0.5$ Pa. Our power-law fit to number of detected vortices indicate the increased activity of convective vortices during the period of study. We have used UV attenuation data to predict the possibility of about 93 dust devil occurrences. Later we use simulations from the Mars WRF model and evaluate the dust devil activity to understand seasonal dependence of these dust devil formation. A seasonal study of UV drops was not done earlier and our result shows that the frequency of occurrence of dust devils increases as we approach the local summer season.

In chapter 4, we numerically solve the equations of motion for dust particles to determine their velocity inside the dust devil and consequently determine the dust distribution using the continuity equation. We consider an initial wind profile, which is dependent on the circulation strength of the vortex (Γ) and viscosity of the air (ν). Our simulations indicate a maximum concentration of ~1400 cm⁻³ near the surface and at the boundary of the vortex. The larger size particles are lifted to lower heights. From the simulated dust distribution in our vortex, we estimate a dust flux of ~5×10⁻⁵ kgm⁻

 2 s⁻¹, a total optical depth of 0.2 which are consistent with observations. Our calculations can provide useful inputs to study the effect of dust devils on boundary layer processes.

In chapter 5, we have determined the magnitude of electric field which can be generated within a dust devil in Martian atmosphere, both in the absence and presence of dust storm. We find that the electric field reaches the breakdown value faster in dust storm scenario as compared to non-dust storm scenario, leading to the fact that lightning can be observed in dust devils when the atmosphere will have dust in it. The charge relaxation will lead to a lower charge buildup in the dust devils thus reducing the strength of the electric field within it. A vertical exponential distribution of charged particles will lead to development of lower electric field than the dipolar configuration of charged particles.

6.2 Scope for future research

In this thesis we have studied about the characteristics of dust devils in Martian atmosphere in steady state by using mean physical conditions of the Martian atmosphere. We did not incorporate the formation and decay of the dust devils in our study and also did not consider their spatio-temporal variability on Mars. Such studies become important if one accounts for the effect of planetary surface forcings on the formation of dust devils. The local topography and solar radiation play a role in the formation of vortices. It is generally understood that during local summer season the number of vortex formations increases, as is discussed in chapter 3 of this thesis. To study in detail these effects on dust devils, we would like to extend the research in this direction and work with a three-dimensional, high resolution model to understand vortex generations. These are known as Large-Eddy Simulations (LES) in which the equations of motion for the atmospheric fluid are integrated on a grid, with a resolution of tens of meters to a few kilometers. This fine resolution resolves the turbulent eddies that account for most of the energy transport within the convective boundary layer. The Martian atmospheric circulations at such scales is highly turbulent owing to high thermal contrasts, short radiative timescales, low atmospheric density and steep topographical gradients [276, 338]. As already described in chapter 3 of this thesis, location plays a major role in the convective vortex formation. Apart from topographical effect, the solar radiation received by that area and the pressure and density of atmosphere too plays a role in vortex formation. We will use this 3D-model to study the effect of surface forcings on the generation of convective vortices on Mars, and their spatio-temporal evolution. In the thesis, we have also determined the dust distribution within a steady state dust devil. In conjunction with a 3D model, this estimate will be useful in determining the total dust load in the atmosphere at various locations and seasons on Mars. This is important, as dust lifting is still parametrized in global 3D models.

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List of Publications

- Uttam, S., Singh, D., & Sheel, V. (2020). Tangential winds of a vortex system in a planetary surface layer. *Journal of Earth System Science*, 129(1), 2. https://doi.org/10.1007/s12040-019-1268-5
- Uttam, S., Sheel, V., Singh, D., Newman, C.E., & Lemmon, M.T., "Meteorological observations of convective vortices at Gale crater, Mars during MY33". Under Review (Planetary and Space Science)
- Sheel, V., Uttam, S., & Mishra, S.K. "Numerical simulation of dust lifting within a steady state dust devil". Under Review (Journal of Geophysical Research)

Presentations in Conference/Symposium

- Shefali Uttam and Varun Sheel, "Dust lifting mechanisms on Mars" presented in Brain Storming Session on Vision and Explorations for Planetary Sciences in Decades 2020-2060 – 2017 held at Physical Research Laboratory, India during 8th – 10th November 2017 [*Poster presentation*].
- Shefali Uttam, Varun Sheel, Deepak Singh, Mark T. Lemmon, "Characteristics of convective vortices and dust devils identified using MSL data" presented in National Space Science Symposium (NSSS) – 2019 held at Savitribai Phule

Pune University (SPPU), Pune during 29th – 31st January 2019 [Poster presentation].

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- Shefali Uttam, "Tangential winds of a vortex system in a planetary surface layer" presented in Indian Planetary Science Conference (IPSC) 2020 held at Physical Research Laboratory, India during 19th 21st February 2020 [Oral presentation].

Publication attached with this thesis

 Uttam, S., Singh, D., & Sheel, V. (2020). Tangential winds of a vortex system in a planetary surface layer. *Journal of Earth System Science*, 129(1), 2. <u>https://doi.org/10.1007/s12040-019-1268-5</u>

Publication attached with this thesis