

Analyzing Mars' Polar Anomalies: Computer Vision Techniques for Seasonal Changes and Polar Dynamics

Pruthviraj (Pruthvi) Acharya

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Department of Earth and Space Science, York University, Toronto, Ontario

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Abstract

This dissertation offers a detailed analysis of seasonal ice cap dynamics and surface anomalies on Mars, utilizing autonomous tracking techniques with polar stereographic images from the Mars Color Imager (MARCI) spanning multiple Mars Years (MY).

This dissertation investigates the recession of the Northern Polar Seasonal Cap (NPSC) from MY 29 to MY 35. Employing Python for automation, this analysis tracks the recession with high temporal fidelity, uncovering intraseasonal variability in the recession rate in addition to significant interannual variability. This variability is coincident with specific events influenced by off-polar winds and Global Dust Storm (GDS) events. The chapter notably examines the divergent effects of GDS events on the size of the NPSC and its recession rates, emphasizing the influence of storm timing and duration.

Additionally, the dissertation explores the recession of the Southern Polar Seasonal Cap (SPSC) from MY 28 to MY 31, characterized by significant discontinuity and variability in recession rates. This study highlights the impacts of the GDS events on SPSC dynamics, demonstrating accelerated sublimation rates and a reduction in cap size after the storm onset.

Finally, the dissertation delves into seasonal phenomena known as Cold and Bright Anomalies (CABAs) and Warm and Dark Anomalies (WADAs) on the North Polar Residual Cap (NPRC). Extensive analysis from MY 29 to MY 35 examines the evolution of temperature and albedo at anomaly sites, revealing a strong correlation with local topography and atmospheric conditions, including katabatic winds and transient eddies.

Collectively, this dissertation provides a nuanced understanding of Martian polar dynamics,

offering insights into the interactions between atmospheric phenomena and surface conditions. The adoption of automated tracking technologies significantly enhances the precision and efficiency of these analyses, contributing to our broader understanding of Martian climatology and its seasonal cycles.

Dedication

Dedicated to my family, Mammi Pappa, and Trisha

Acknowledgments

I extend my deepest appreciation to Professor Isaac Smith, my PhD advisor, for his unwavering support and guidance throughout my undergraduate and graduate research. His insights and expertise were indispensable to the completion of this dissertation, and I am particularly grateful for his guidance in steering my work in the right direction.

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To my friends, my Canadian friends and my international friends, you have been essential throughout this entire journey. I am also grateful to all of my labmates for their companionship and support for this research.

Finally, I owe a profound debt of gratitude to my family—my parents, Mammi and Pappa, and my little sister, Trisha. Their endless support, boundless patience, and belief in my abilities have been my anchors throughout this entire PhD journey.

This journey has truly been a collaborative effort, and I am thankful to everyone who has played a part, however small, in helping me reach this milestone.

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List of Acronyms

- CABA: Cold and Bright Anomalies
- CTX: Context Camera
- CRISM: Compact Reconnaissance Imaging Spectrometer for Mars
- GDS: Global Dust Storm
- HiRISE: High Resolution Imaging Science Experiment
- L_s : Solar Longitude
- MARCI: Mars Color Imager
- MCS: Mars Climate Sounder
- MM: Mountains of Mitchel
- MRO: Mars Reconnaissance Orbiter
- MOLA: Mars Orbiter Laser Altimeter
- MY: Mars Year
- NPRC: North Polar Residual Cap
- NPSC: North Polar Seasonal Cap
- NPLD: North Polar Layered Deposits

- PLD: Polar Layered Deposits
- PRC: Polar Residual Cap
- PSC: Polar Seasonal Cap
- SHARAD: Shallow Radar
- SPLD: South Polar Layered Deposits
- SPRC: South Polar Residual Cap
- SPSC: South Polar Seasonal Cap
- THEMIS: Thermal Emission Imaging System
- WADA: Warm and Dark Anomalies

Preface

Sections from the following chapters of this dissertation have been adapted from previously published peer-reviewed papers or papers in review:

Chapter 2: Acharya, P. J., Smith, I. B., & Calvin, W. M. (2023). Tracking the northern seasonal cap retreat of mars using computer vision. *Icarus*, 390, 115295. <https://doi.org/10.1016/j.icarus.2022.115295>

Chapter 3: Acharya, P. J., Smith, I. B., & Calvin, W. M. (2024). Tracking the south polar seasonal cap retreat of Mars using computer vision. <https://doi.org/10.1016/j.icarus.2024.116104>

Chapter 1

Introduction

1.1 Overview of Mars

1.1.1 Physical Characteristics of Mars

Mars, often compared to Earth, displays a distinct geological and geographical landscape. Mars has a surface area of 144.4 million square kilometers — comparable to Earth's total landmass (Catling 2014). Mars also holds about 15% of Earth's mass and 11% of its volume, resulting in a surface gravity of approximately 38% that of Earth, or 3.7 meters per seconds⁻² (Catling 2014) (Figure 1.1)1. The Martian surface, reminiscent of Earth's deserts, is characterized by a red-orange hue primarily due to iron oxide (Catling 2014). A Martian day lasts 24.6 hours, just 39 minutes longer than Earth's 24-hour day, and a Martian year is 1.9 Earth years (Catling 2014). Mars orbits the Sun at 1.5 AU (222.96 million kilometers) compared to Earth's 1 AU and has an eccentricity of 0.093 compared to Earth's 0.0167 (Catling 2014). The larger Martian orbit means that Mars receives total solar irradiance, a measurement of total incoming solar energy, of approximately 586.2 watts per square meter compared to Earth at 1361.0 watts per square meter, resulting in a lower energy budget (Catling 2014). Mars's axial tilt of 25.1° leads to pronounced seasonal changes, similar to Earth's (Catling 2014; Haberle 2015). Like Earth, Mars undergoes Milankovitch cycles—variations in orbital eccentricity, axial tilt, and precession (Laskar et al. 2004). These cycles can shift Mars'

tilt between 15° and 46° , which could induce more extreme seasonal variability than currently observed (Laskar et al. 2004). The Martian terrain features ancient river systems, mountain ranges, and canyons primarily shaped by wind, water and glacier erosion, suggesting a past climate that could have supported liquid water and potentially life (Pieri 1980; Wordsworth 2016).

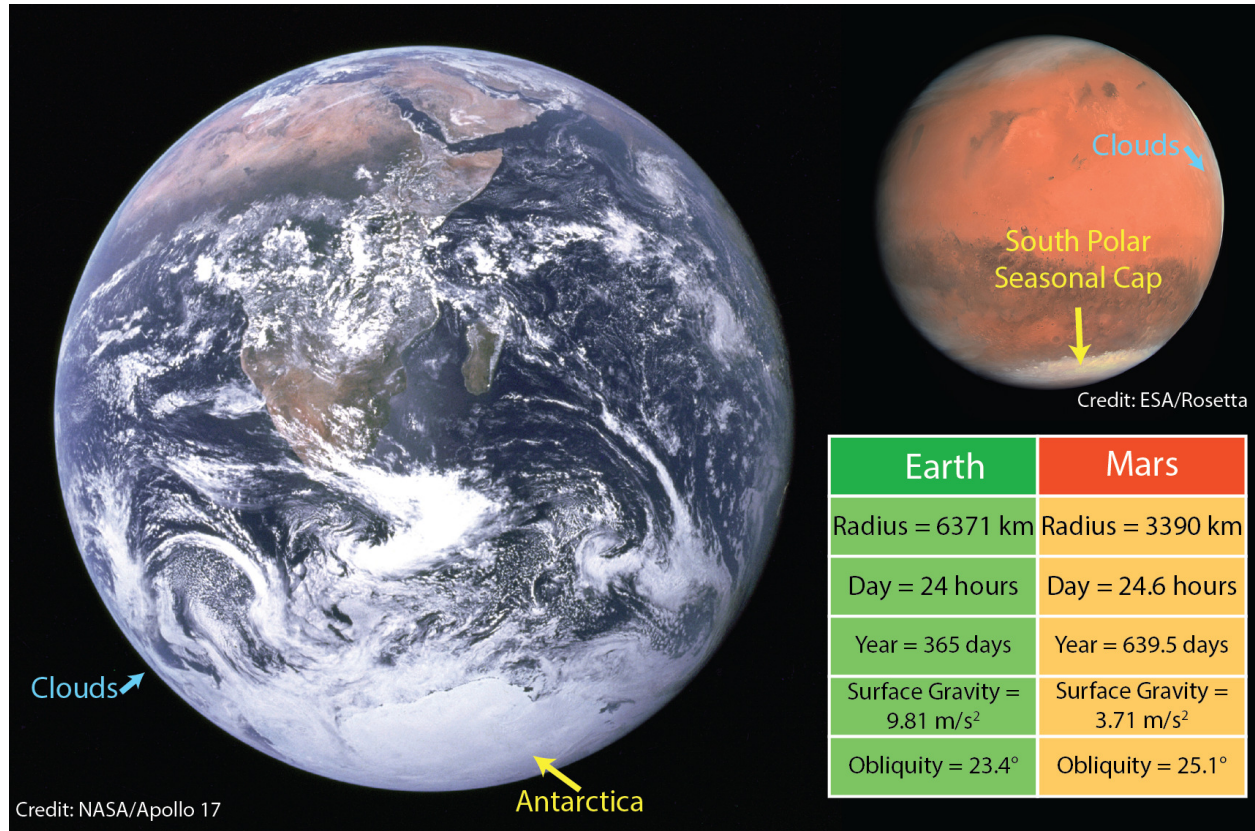


Figure 1.1: General overview between Earth and Mars (to scale). The left image is the famous "Blue Marble" image taken by the Apollo 17 crew in 1972. The right image is a true colour image taken by the OSIRIS instrument onboard the Rosetta spacecraft in 2007. Both planets have similar ice-capped poles (Antarctica vs South Polar Seasonal Cap) and atmospheric activity (Clouds).

1.1.2 Overview of the Martian Atmosphere

The atmosphere of Mars consists predominantly of carbon dioxide (CO_2), which constitutes 96% of its volume. It also contains 2.85% nitrogen and 0.03% water, a stark contrast to Earth's nitrogen and oxygen-rich atmosphere (Trainer et al. 2019). Trace amounts of water vapour (0.03%), oxygen (0.13%), ozone (0.03 ppm), and various noble gases also permeate the Martian atmosphere (Haberle 2015; Franz et al. 2017; Trainer et al. 2019). Mars features a much colder and thinner

atmosphere compared to Earth. The highest measured air density on Mars reaches about 20 grams per cubic meter, only 2% of Earth's surface air density (Barth 1974; Savijärvi and Kauhanen 2008). The average surface pressure on Mars hovers around seven millibars, roughly 1% of Earth's, with seasonal fluctuations causing it to vary between six and eight millibars (Moroz 1998; Trainer et al. 2019). The pressure fluctuations are highly dependent on location and elevation (Trainer et al. 2019).

Similar to Earth, Mars' atmosphere is organized into distinct layers, each analyzed through temperature and pressure measurements at various altitudes to create vertical profiles (Haider 2023). Mars has four main atmospheric layers compared to Earth's five, including the troposphere, mesosphere, thermosphere, and exosphere (Figure 1.2). Mars does not possess a stable stratosphere as the Martian atmosphere lacks significant ultraviolet absorbers such as ozone in Earth's atmosphere and methane in Titan's atmosphere. When ultraviolet light from the Sun is absorbed by the absorbers, the molecule is excited and increases the energy state, which is then released as heat. This process heats the surrounding air, leading to an increase in temperature with altitude in the stratosphere (Figure 1.2). This heating creates a stable layer known as the stratosphere. Nonetheless, a seasonally occurring stratosphere is occasionally observed over the southern polar region of Mars (McCleese et al. 2008). Each Martian atmospheric layer is significantly larger than its counterpart on Earth, with the Martian troposphere extending up to 40 kilometers from the surface—substantially higher than Earth's troposphere, which reaches about 11 kilometers (Haberle 2015).

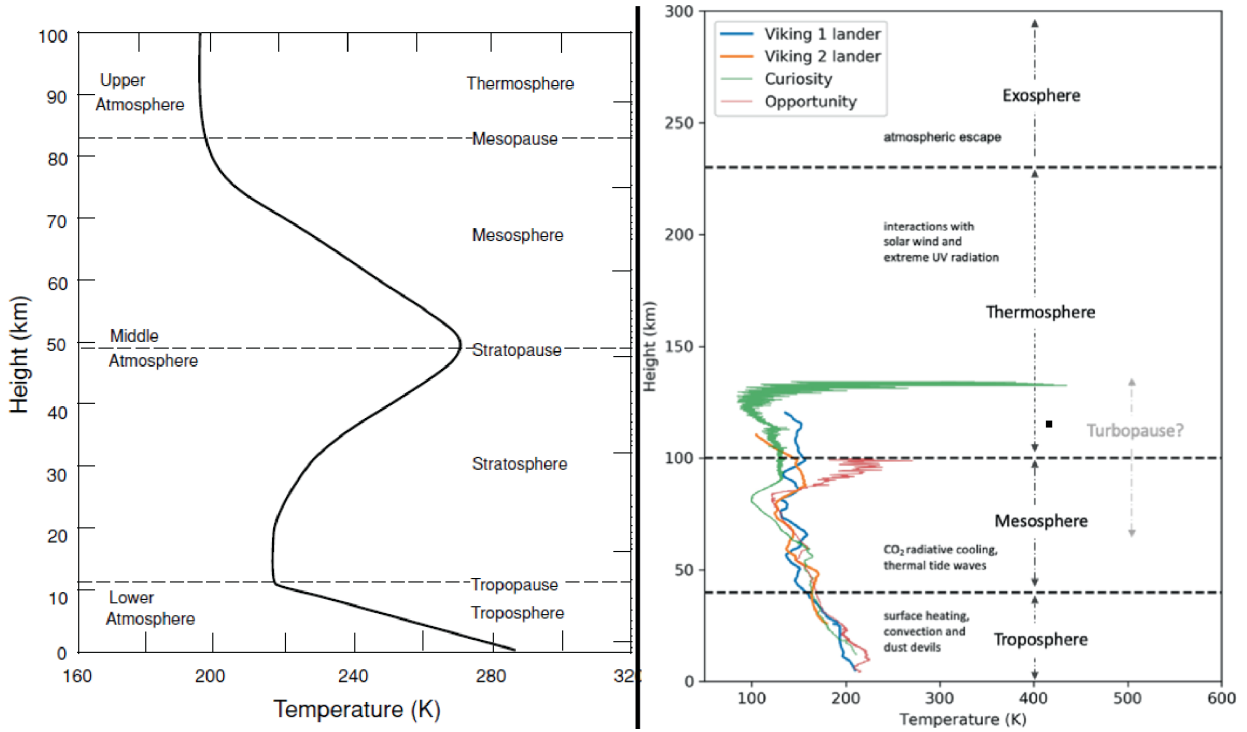


Figure 1.2: Left shows the terrestrial temperature profile from ? and right shows the Martian temperature profile measured by various landers during decent (Haider 2023). The vertical axis represents the altitude in kilometers (km) and the horizontal axis represents the measured temperature in kelvin (K). Earth's troposphere extends to about 11 kilometers followed by an increase in the temperature due to the stratosphere. The Martian troposphere extends up to 40 kilometers and lacks a stratosphere due to the absence of UV absorbers. The Martian atmospheric layers are significantly larger than their terrestrial counterparts.

The Martian troposphere, the site of most of the planet's weather and dynamic atmospheric phenomena, is heavily influenced by dust (Haberle 2015). Airborne dust plays a critical role in the planet's climate by absorbing and re-emitting solar radiation, thereby significantly affecting Mars' radiative balance (Leovy et al. 1973; Pollack et al. 1979). Martian winds, which can exceed speeds of 30 meters per second, are capable of lifting and transporting dust particles across the planet under favourable conditions (Leovy et al. 1973; Kok et al. 2012). Airborne dust plays a more critical role than surface dust due to its significant impact on atmospheric dynamics and the Martian weather system. Airborne dust is lifted by Martian winds and distributed globally, influencing temperature and playing a key role in precipitation (Plesa et al. 2016; Gary-Bicas et al. 2020). Moreover, the presence of airborne dust affects the accuracy of optical scientific measurements and

operational capabilities of surface missions. Dust activity on Mars shows considerable seasonal and spatial variation, reaching its peak during the southern spring and summer. Occasionally, this activity escalates into Global Dust Storm (GDS) events, which can envelop the entire planet (Sánchez-Lavega et al. 2019; Wolkenberg et al. 2020) (Figure 1.3).

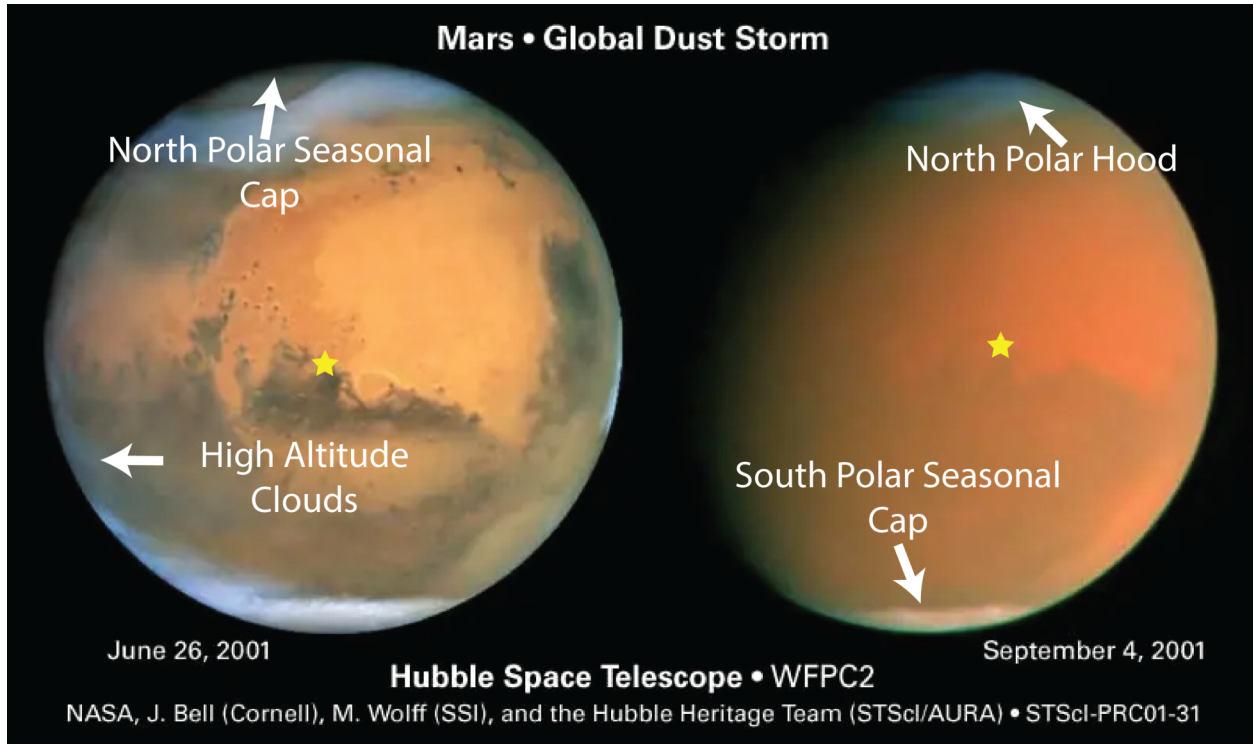


Figure 1.3: Hubble Observations during the 2001 (MY 25) Global Dust Storm (GDS) Event. The star roughly marks the Syrtis Major Planum. The left image shows Mars during near north fall ($L_s = 180^\circ$; Martian date convention explained in Section 1.1.3) with prominent cloud features and the North Polar Seasonal Cap (NPSC) and the South Polar Seasonal Cap (SPSC). The right image shows Mars during the GDS near the winter solstice ($L_s = 226^\circ$) with the storm covering the entire planet.

1.1.3 Martian Timekeeping

To characterize the seasonal processes on Mars, this dissertation uses the solar longitude (L_s) convention. This convention segments the Martian orbit into degrees from 0° to 360° , representing Mars' position along its orbit (Figure 1.4). The phases in the Martian orbit are: $L_s = 0^\circ/360^\circ$ marks the onset of spring in the northern hemisphere, $L_s = 90^\circ$ coincides with the summer solstice, $L_s = 180^\circ$ signifies autumn equinox, and $L_s = 270^\circ$ aligns with the winter solstice - the opposite

season in the southern hemisphere (Table 1.1). Additionally, the Mars Year (MY) system is used to chronologically categorize Martian years, starting with MY 1 in 1955 (Piqueux et al. 2015a).

Table 1.1: The Martian L_s conversion

L_s date [°]	Season in the North Polar Region	Season in the South Polar Region
0°/360°	Spring Equinox	Autumn Equinox
90°	Summer Solstice	Winter Solstice
180°	Autumn Equinox	Spring Equinox
270°	Winter Solstice	Summer Solstice

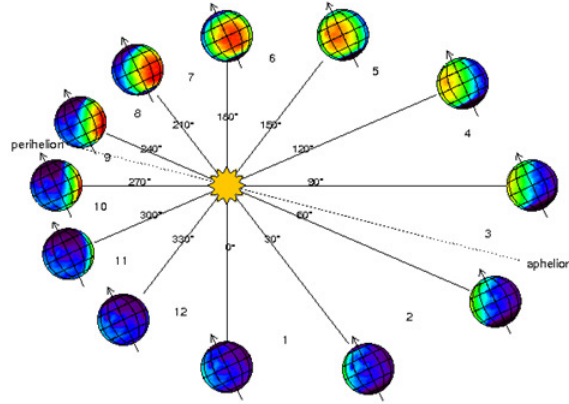


Figure 1.4: Season on Mars and solar longitude (L_s). The L_s angles denote the Sun-Mars angle where $L_s = 0$ is the northern spring equinox. The figure also shows the eccentricity of Mars, which allows for southern winters to be colder and longer. (Forget 2008)

1.2 Overview of the Polar Regions of Mars

The northern and southern polar regions of Mars, each extending about 40° latitude from their respective geographic poles, stand as some of the most dynamic areas on the Martian surface in terms of annual variations and seasonal activities. These regions exhibit large regional dust storms exceeding millions of square kilometers to localized surface clouds less than 10 kilometers long. These regions host seasonal CO₂ ice, permanent residual ice caps, and subsurface layered deposits. Moreover, unique features of the northern polar region distinguish it from the southern, indicating differences in their geological histories and seasonal dynamics.

1.2.1 General Overview of Important Polar Features

The Polar Seasonal Caps (PSCs), some of the most extensive surface features on Mars, are visible with ground-based telescopes (James et al. 1987; Parker et al. 1999) (Figures 1.3, 1.5). At their peak, these caps cover approximately 20 million square kilometers (Brown et al. 2010; Piqueux et al. 2015b), representing about 14% of the Martian surface or twice the size of Canada. The PSCs are primarily composed of H₂O (water) and CO₂ ice, with latitudinal and temporal variability in the thickness of the cap, with the thickest ice present near the center of the PSCs during polar winter (Aharonson et al. 2004; Giuranna et al. 2007; Prettyman et al. 2009). Notably, parts of the CO₂ cap transform into slab CO₂ ice, transparent to solar radiation, affecting the interaction with solar energy (Titus et al. 2001; Kieffer et al. 2006). The dynamics of these caps significantly influence the Martian climate and landscape, topics further explored in Chapter 1.2.3.

Additionally, the Martian polar hood, a prominent cloud formation visible from both orbital and ground-based telescopes, envelopes the PSCs (Parker et al. 1999; Benson et al. 2011) (Figure 1.3). This cloud zone typically emerges in the polar winters and can extend into the spring, primarily driven by the polar vortex—a circumpolar wind that regulates the movement of aerosols, including dust, within the polar regions (Benson et al. 2011; Streeter et al. 2021). The North Polar Hood is particularly noted for its water-ice clouds that persist from early autumn ($L_s = 150^\circ$) to spring ($L_s = 30^\circ$), consistently shrouding the seasonal cap (Benson et al. 2011).

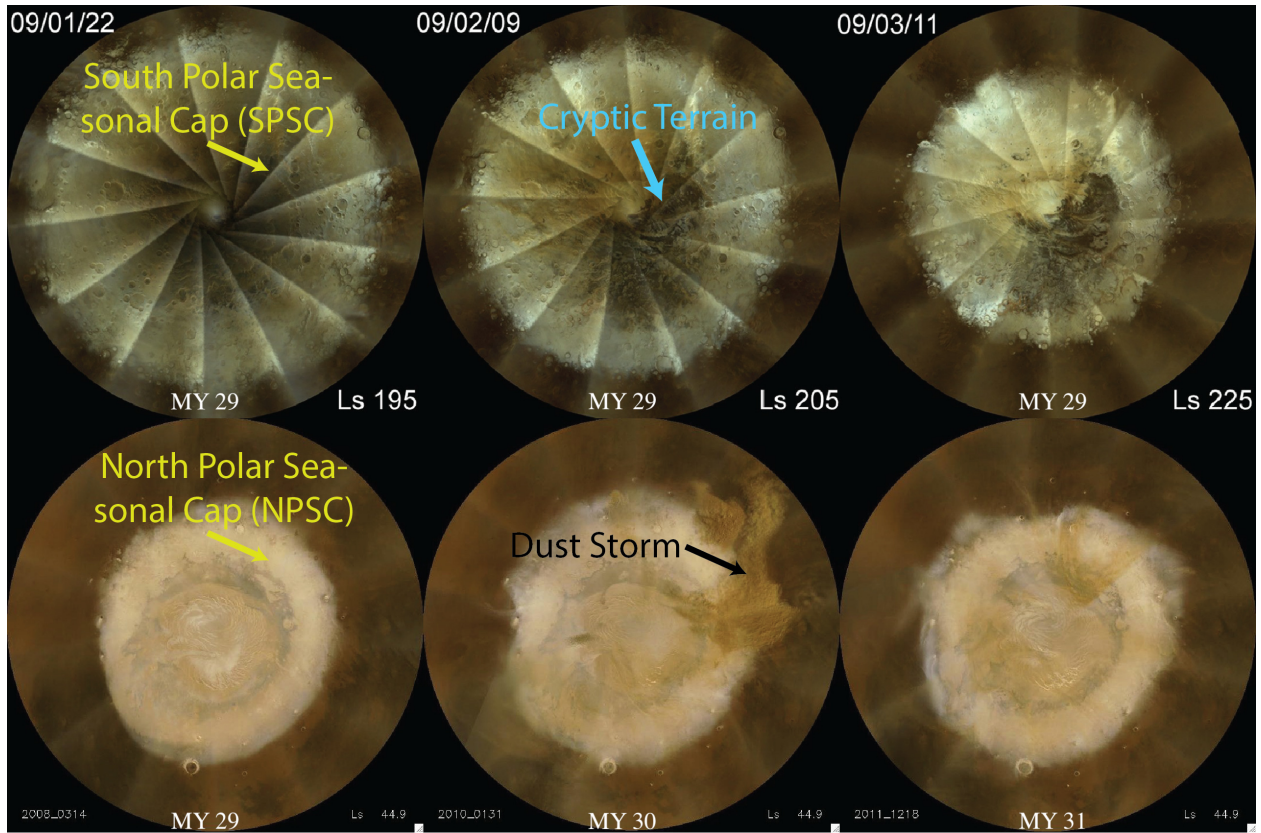


Figure 1.5: Polar mosaics from the Mars Color Imager (MARCI) onboard the Mars Reconnaissance Orbiter monitoring the recession of the PSCs for multiple MYs. Top row: Figure 2 from Calvin et al. (2017) showing the retreat of the South Polar Seasonal Cap (SPSC) during southern spring. Bottom row: Figure 2 from Calvin et al. (2015) showing the NPSC at spring $L_s = 44.9^\circ$ for multiple MYs. Left to right: MY 29, MY 30, MY 31. Yellow Arrow marks the North Polar Seasonal Cap (NPSC) and the South Polar Seasonal Cap (SPSC), the black arrow marks a large regional dust storm during MY 30, and the blue arrow marks the Cryptic Terrain (darker than the SPSC but still made up of CO_2 ice).

The Polar Residual Caps (PRCs) of Mars, composed of CO_2 and water-ice, withstand the polar summer surface temperatures, which can ascend to as high as 300 K (Giuranna et al. 2007; Brown et al. 2016) (Figure 1.6). Radar investigations reveal that this residual cap overlays the Polar Layered Deposits (PLD), consisting of stratified layers of water-ice interspersed with dust (Fishbaugh et al. 2010). The variable morphology of these layers, shaped by interactions between the lower troposphere and the surface, indicates long-term climatic shifts on Mars, particularly with changes in the planet's obliquity and orbital eccentricity (Laskar et al. 2004; Smith et al. 2016; Emmett et al. 2020) (Figure 1.6). Studies and models suggest that periods of low obliquity correlate with increased rates of net deposition within these layers, resulting in more pronounced formations

(Laskar et al. 2004). Notably, within the PLD, there are spiral-shaped depressions, similar to deep troughs, reaching depths of up to 500 meters in some locations. These troughs have migrated over geological timescales, undergoing continuous erosion and deposition cycles (Smith and Holt 2010; Smith et al. 2013) (Figure 1.6). The dynamic nature of these troughs introduces additional complexity to our understanding of the Martian polar regions.

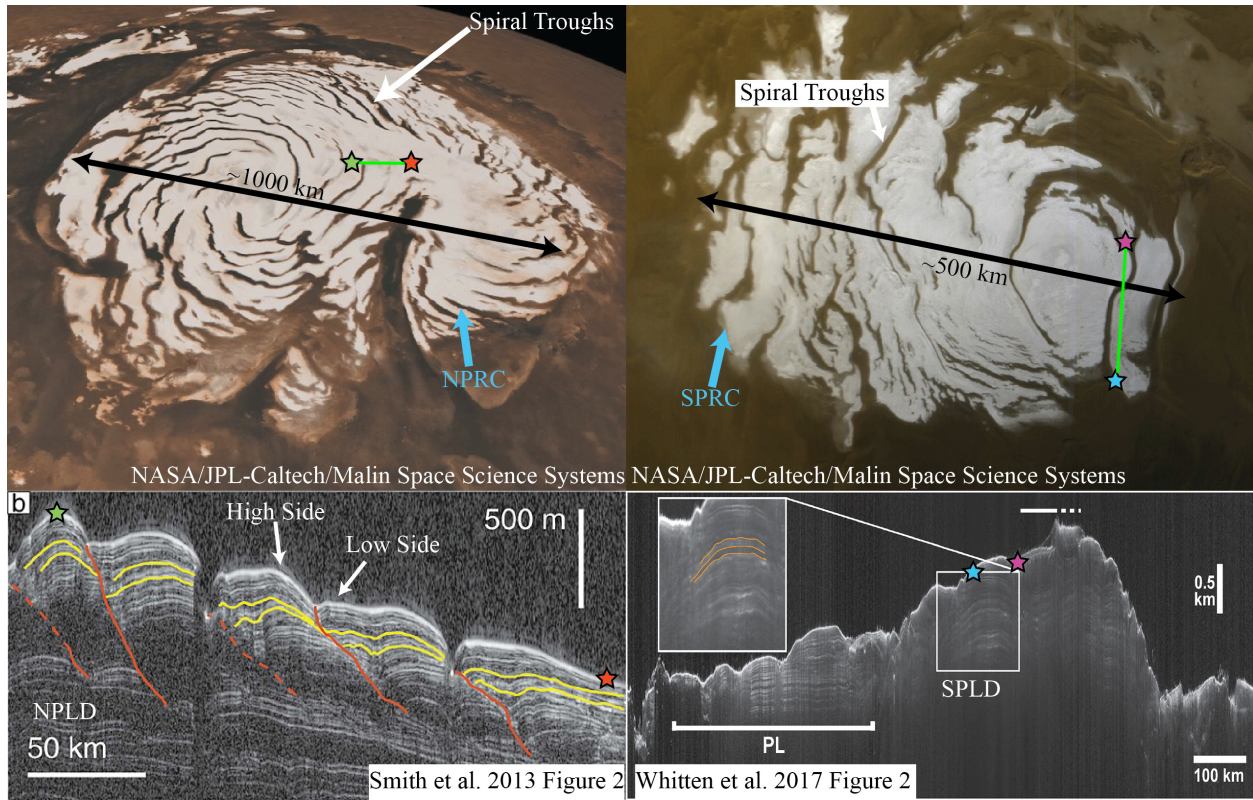


Figure 1.6: The top row shows the North Polar Residual Cap (NPRC; Top Left), and the South Polar Residual Cap (SPRC; Top Right) as seen from visual band images with the spiral troughs as dark dust bands within the bright ice (White Arrow marks one such trough). The coloured stars mark the endpoints of the radargram track (Green Arrow) seen in the bottom row. The bottom row shows the North Polar Layered Deposits (NPLD; Bottom Left), and the South Polar Layered Deposits (SPLD; Bottom Right) radargram (Green Line in Top Row) and the Polar Layers (PL) present underneath the surface. The NPLD radargram is Figure 2 from Smith et al. (2013) and the SPLD radargram is Figure 2 from Whitten et al. (2017), showing the dichotomy between the two subsurface layers. The coloured stars mark the endpoints of the radargram seen in the respective visual images.

1.2.2 General Overview of the Seasonal Polar Processes occurring within a Mars Year

PSCs start to form during polar autumn at $L_s = 180^\circ$ in the north polar region and $L_s = 0^\circ$ in the south polar region—as the surface temperature drops below the CO_2 condensation point of approximately 150 K (Leighton and Murray 1966) (Figure 1.7). The PSCs form due to direct condensation, when the surface is cold enough to condense CO_2 and water from the atmosphere, and precipitation such as snowfall. The PSCs continue to expand until the peak of winter, at $L_s = 270^\circ$ in the north and $L_s = 90^\circ$ in the south (Piqueux et al. 2015b). Subsequently, until late spring— $L_s = 80^\circ$ in the north and $L_s = 200^\circ$ in the south—the PSCs begin to retreat, starting at the margins of the cap, as the surface temperatures within the region rise above the CO_2 condensation point (Piqueux et al. 2015b) (Figure 1.5).

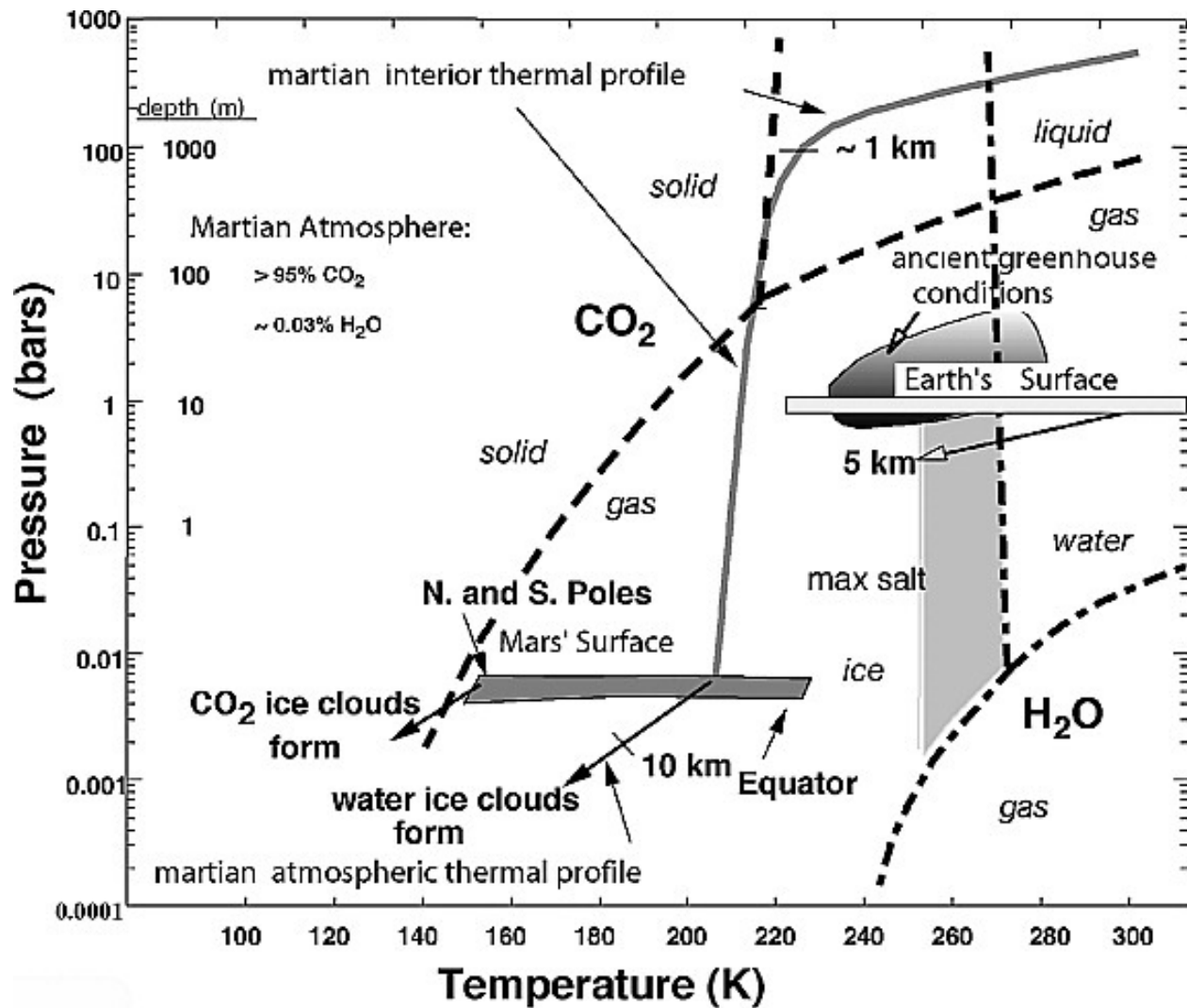


Figure 1.7: Figure 1 from Longhi (2006) showing the phase diagram of pure CO₂ (Dashed Line) and pure water (Dashed-Dot Line). The dark box in the lower left corner marks the pressure and temperature ranges found on the surface of Mars. The surface pressure ranges from 0.01 bar to 0.005 bars and the surface temperature ranges from 140 K to 220 K. During polar winters, the surface temperature can reach the solid CO₂ line at 150 K and surface pressure of 6 millibars, allowing for the polar caps to form. The seasonal caps quickly sublimate away as the surface temperature reaches above that point. Liquid water is not possible as the surface pressure and surface temperature are too low.

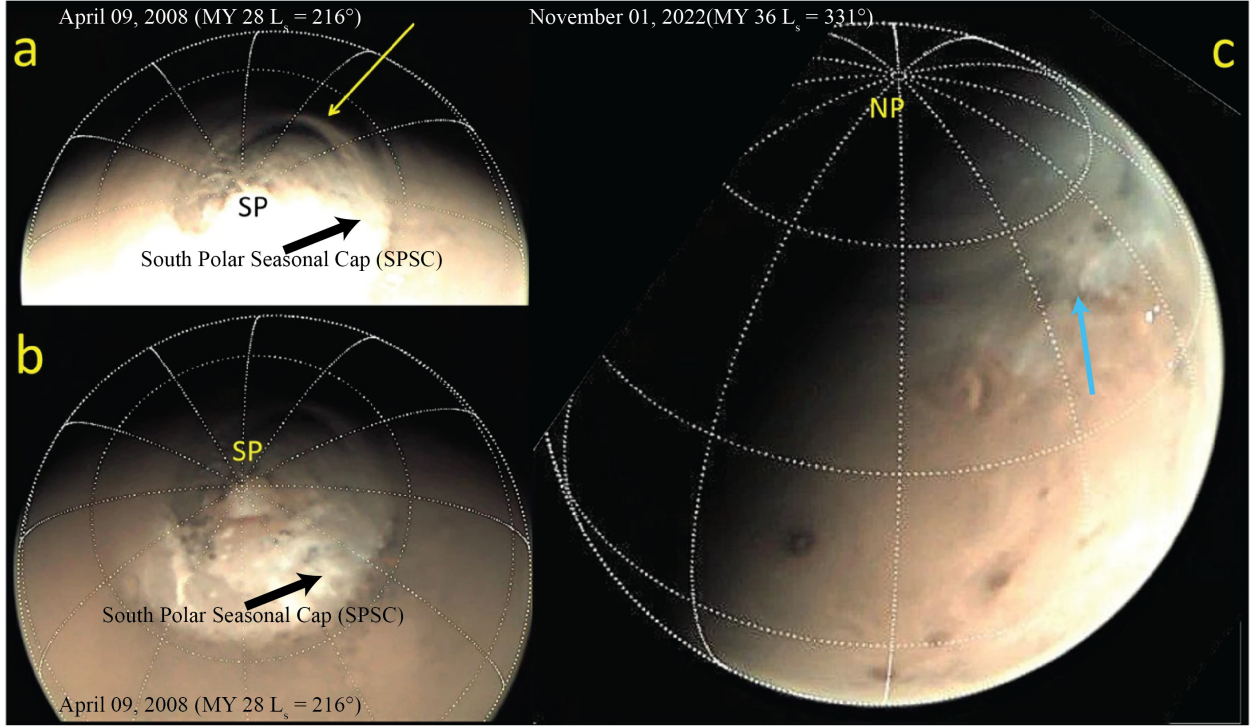


Figure 1.8: Figure 6 from Sánchez-Lavega et al. (2024) shows various seasonal activities captured by the Visual Monitoring Camera onboard the Mars Express. (a) and (b) show the MY 28 (2018) GDS event (yellow arrow) development at $L_s = 216^\circ$ and (c) shows the NPH clouds at MY 36 $L_s = 331^\circ$ (blue arrow).

From the spring equinox to late spring, at $L_s = 80^\circ$ in the northern polar region and $L_s = 200^\circ$ in the southern polar region, the polar regions experience intense and dynamic seasonal activities. During this period, large regional dust storms, capable of extending thousands of kilometers, frequently originate near the retreating edge of the PSC and spread to mid-latitude regions (Leovy et al. 1973; James et al. 1996; Wang and Richardson 2015) (Figures 1.5, 1.8, 1.9). Accompanying these vast dust storms are water-ice clouds, significantly smaller in scale, forming near the retreating edges of the PSC. These clouds span only hundreds of kilometers and are believed to form as water sublimates from the edge and subsequently re-deposits onto the PSC (Appéré et al. 2011; Calvin et al. 2015; 2017) (Figures 1.5, 1.8, 1.9).

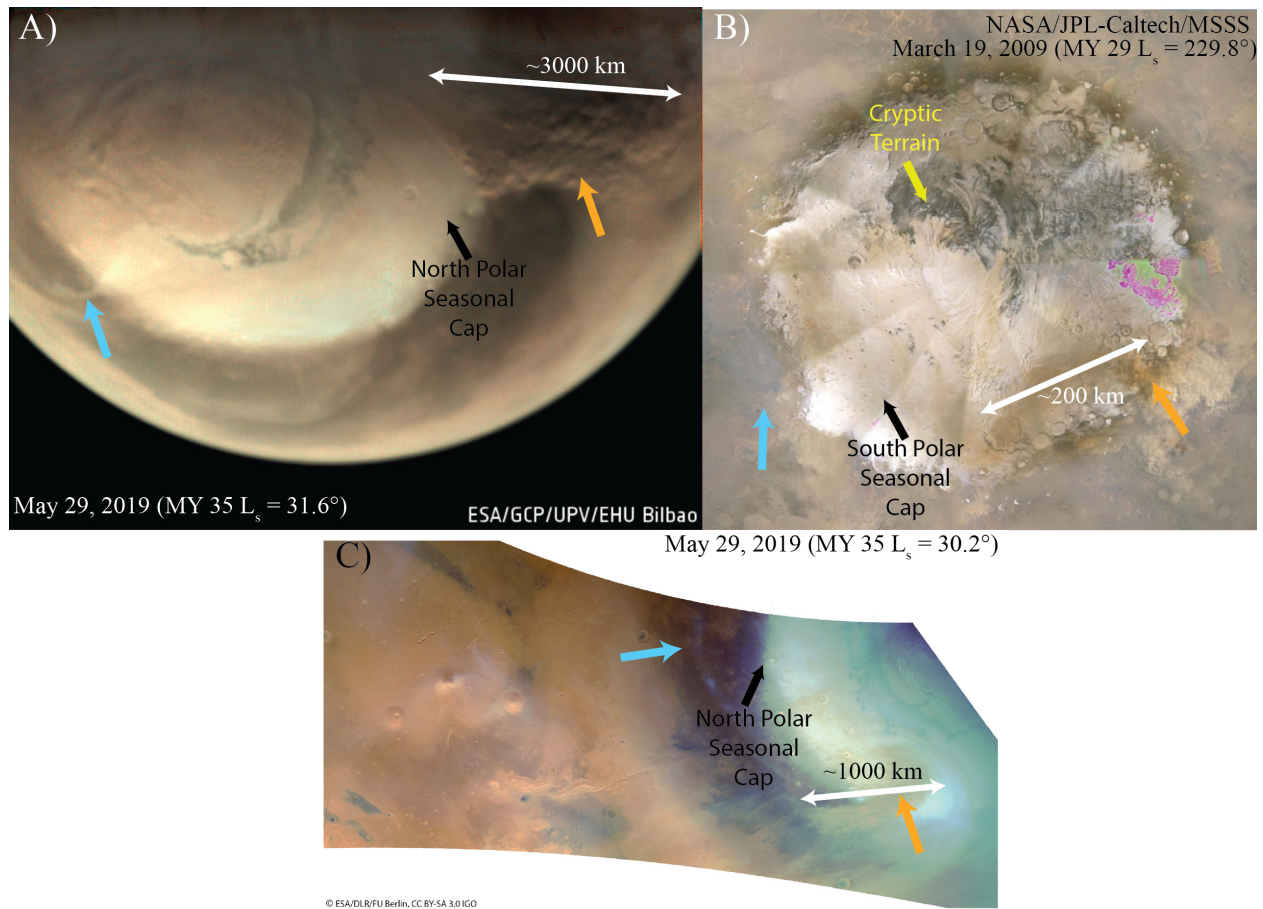


Figure 1.9: Figures showing various seasonal activities occurring during the Martian polar regions. The orange arrows mark an example of a large regional dust storm. The blue arrows mark an example of the water-ice clouds seen at the margins of the seasonal caps. A) A large dust storm was seen near the margins of the North Polar Seasonal Cap (NPSC) by the Visual Monitoring Camera onboard the Mars Express. B) A relatively large regional dust storm was seen near the margins of the South Polar Seasonal Cap (SPSC) by the Mars Color Imager (MARCI) on board the Mars Reconnaissance Orbiter. C) A larger spiral storm was seen at the margins of the NPSC by the High Resolution Stereo Camera (HRSC) imager on Mars Express.

As late spring approaches, about $10^\circ L_s$ dates before the summer solstice, the PSC predominantly covers the PLD, marking a shift in seasonal activities (Appéré et al. 2011; Smith et al. 2013; 2015; Smith and Spiga 2018). During this period, the lower troposphere exhibits strong interactions, leading to the formation of surface streaks driven by katabatic winds, which are near-surface winds located less than 2 kilometers above the ground (Howard 2000; Spiga and Smith 2018). These winds originate from cold, dense air at higher elevations moving downwards, analogous to a ball rolling downhill, and are subsequently steered by the Coriolis Force as they move equatorward

(Howard 2000; Smith et al. 2013). The influence of the "sea-breeze" effect, characterized by the sharp thermal gradient between the seasonal ice and the adjacent warmer terrain, including residual ice, further amplifies these winds (Smith et al. 2015). Reaching speeds exceeding 20 meters per second, the katabatic winds not only transfer momentum and energy to the surface, enhancing local sublimation rates but also facilitate the transport of dust and volatiles such as CO₂ and water (Smith et al. 2013; Smith and Spiga 2018; Spiga and Smith 2018). This material is deposited downwind through precipitation or condensation (Smith et al. 2013).

Additionally, these winds interact with the Martian topography, particularly the spiral trough features; they accelerate upon entering a trough and decelerate upon exiting (Smith et al. 2013). This variation in wind speed at the trough base induces changes in flow height, pressure, and temperature, phenomena collectively referred to as a katabatic jump (Smith et al. 2013; Spiga and Smith 2018) (Figure 1.10). At this point, the conditions within the flow allow for the formation of water-ice crystals that appear as near-surface clouds in visual images and enable the precipitation of volatiles such as CO₂ and water (Smith et al. 2013; 2015; Smith and Spiga 2018; Spiga and Smith 2018). The deposition from these clouds plays a pivotal role in the gradual migration of the troughs, thereby significantly contributing to the dynamic reshaping of the Martian surface (Smith et al. 2013).

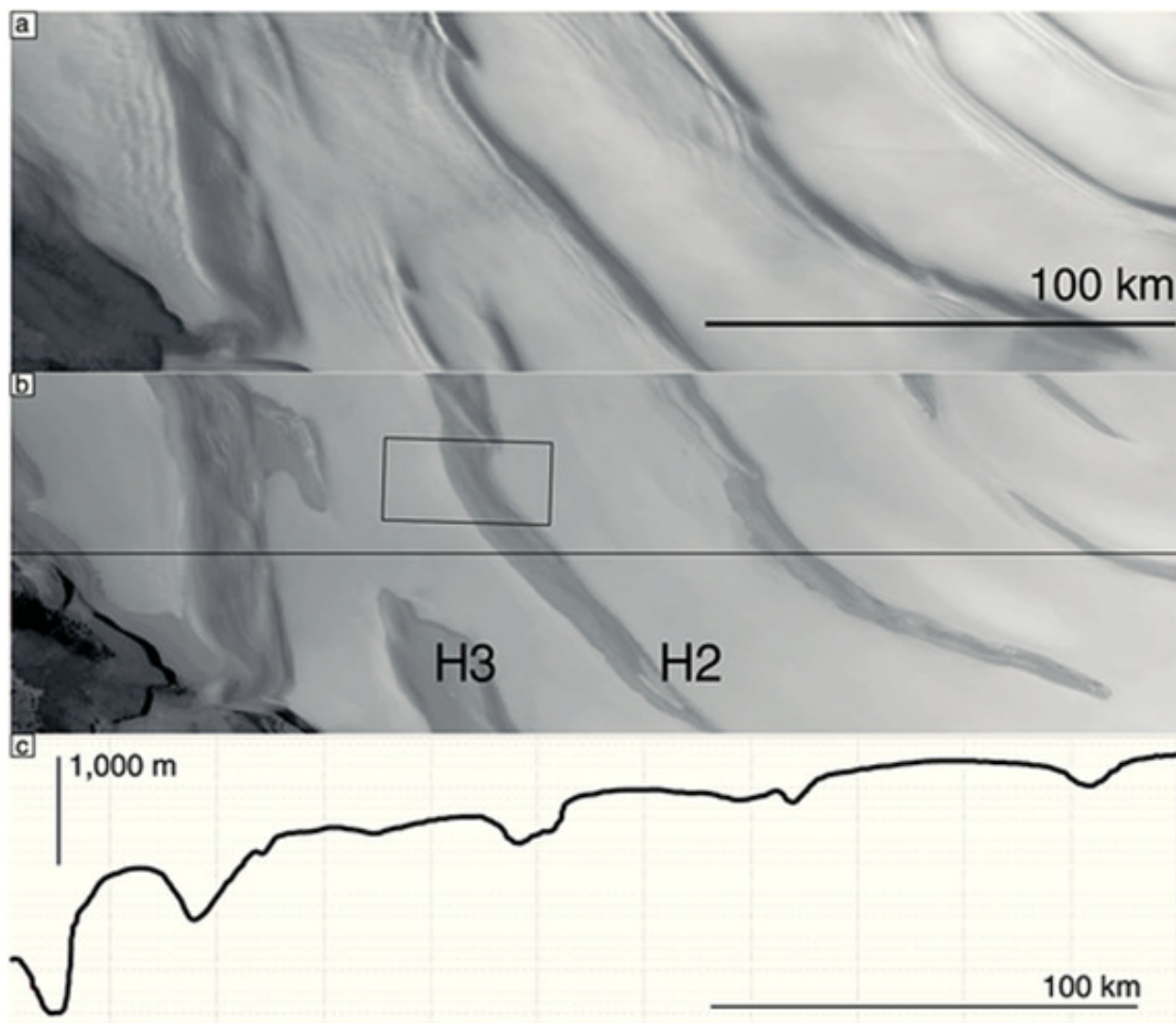


Figure 1.10: Figure 6 from Smith et al. (2013) shows the near-surface clouds associated with katabatic wind flow interacting with the North Polar Layered Deposition that appears during late spring (a) and does not appear during summer (b). (b) horizontal line shows the approximate cross-section corresponding to (c) - surface topography from the Mars Orbiter Laser Altimeter (MOLA).

Following the summer solstice—marked by $L_s = 90^\circ$ in the north and $L_s = 270^\circ$ in the south—there is a noticeable reduction in seasonal activity (Smith et al. 2013; Calvin et al. 2015; 2017; Smith and Spiga 2018; Lutz et al. 2024). During this phase, the seasonal ice has completely sublimated, exposing the water-ice-rich PRC (Giuranna et al. 2007; Brown et al. 2016) (Figure 1.6). As late summer approaches, just before the autumn equinox— $L_s = 180^\circ$ in the north and $L_s = 0^\circ$ in the south—there is a resurgence in atmospheric activity, characterized by the formation of late-season water-ice clouds at the edges of the polar regions (Calvin et al. 2015; 2017; Hinson et al. 2022).

As the autumn equinox nears, these clouds migrate toward the geographic poles, and large spiral dust storms begin to form, moving rapidly across the region within a few hours (Calvin et al. 2015; 2017). Concurrently, the polar hood begins to form, and the early stages of the PSC reemerge as CO₂ starts to condense back onto the surface, marking the seasonal cyclical changes (Benson et al. 2010; 2011; Piqueux et al. 2015b; Alsaeed and Hayne 2022).

1.2.3 Role of the Martian Polar Regions in the Martian Climate

The polar regions of Mars are pivotal in shaping the planet's past, present and future climate and atmospheric dynamics. During the polar winters, as the PSCs grow outwards from the geographic pole, up to 30% of the Martian atmospheric CO₂ and water condenses within these caps, markedly influencing the global atmosphere (Tillman et al. 1993; Forget and Pollack 1996; Kelly et al. 2006). With the start of polar spring and summer, the sublimation of the PSCs reintroduced the previously trapped CO₂ and water back into the atmosphere. This cyclical process of condensation and sublimation drives significant fluctuations in surface pressures across Mars, as evidenced by observations from the Mars Science Laboratory Rover at Gale Crater, located at 5.4° S, and the InSight Lander at Elysium Planitia, situated at 4.5° N (Ordonez-Etxeberria et al. 2018; Martínez et al. 2017) (Figure 1.11). In addition to the surface pressure fluctuations, the margins of the PSCs exhibit intense seasonal activity such as intense regional dust storms and water-ice clouds that highlight the movement of volatiles (CO₂, water, and dust) in and out of the polar region. Moreover, the PRCs act as vital reservoirs of water-ice, crucial to the Martian water cycle (Haberle 2015).

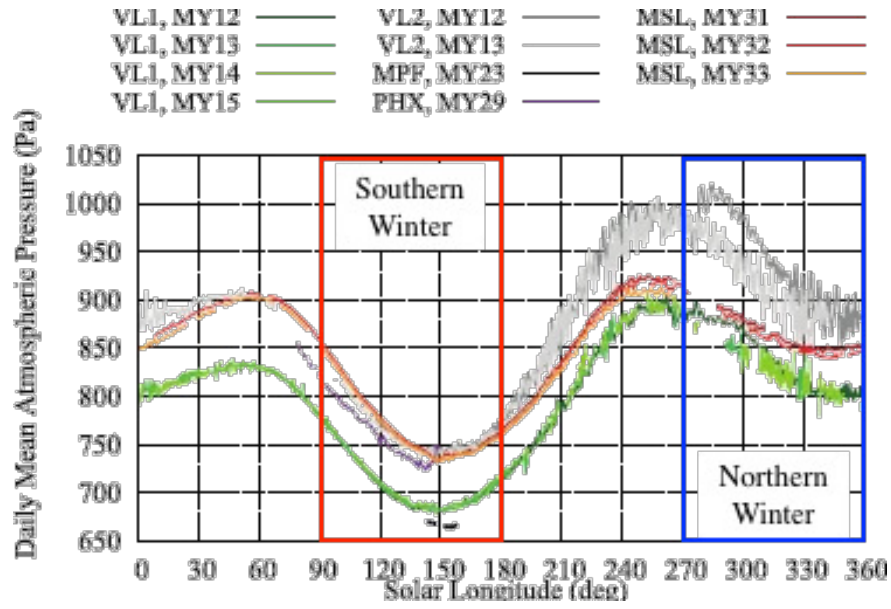


Figure 1.11: Figure 3 from Martínez et al. (2017) showing the seasonal surface pressure variations from various Martian missions. A major decrease in the surface pressure coincided with the formation of the PSCs ($L_s = 90^\circ$ - 150° and $L_s = 250^\circ$ - 340°). Colder and longer southern winter (Red Box) results in a pressure decrease of 30% due to the formation of the South Polar Seasonal Cap (SPSC). In contrast, North Polar Seasonal Cap (NPSC) results in a pressure decrease of 10% (Blue Box).

The polar regions critically influence atmospheric circulation patterns on Mars, primarily through the temperature gradients between the poles and the equator. Thermal models illustrate that these gradients are responsible for generating wind currents that can transport dust across the Martian surface, both towards and away from the poles (François et al. 2014). This interplay between thermal gradients and atmospheric circulation highlights a complex array of factors that shape the Martian climate (Leovy et al. 1973; Pollack et al. 1979; Forget and Pollack 1996).

1.3 Comparison between the North and South Polar Regions of Mars

While the northern and southern polar regions of Mars exhibit similarities in their seasonal processes and surface features, they also present distinct differences that profoundly impact the Martian climate and landscape. Additionally, each region exhibits unique anomalies that further

distinguish the two regions.

1.3.1 Comparison of the topography between the two polar regions

Vastitas Borealis, the northern lowlands/plains is the largest lowland region in the northern latitudes of the planet. This region encircles the north polar region and lies about five thousand kilometers below the average elevation of Mars (Smith et al. 2001). This area is characterized by a sparse distribution of craters and a relatively uniform landscape that predominates across the northern hemisphere (Smith et al. 2001) (Figure 1.12). In contrast, the southern polar region lies about one kilometer above Mars' mean elevation and is characterized by a dense concentration of craters and varying landscape (Smith et al. 2001) (Figure 1.12). Age estimates suggest that the southern highlands are around four billion years old, making them about one billion years older than the northern lowlands (Werner et al. 2011; Rodriguez et al. 2020). A significant geological feature of note is Hellas Planitia, a massive impact crater located near the extremes of the southern polar region. It spans seven thousand kilometers in diameter and reaches a depth of eight kilometers (Smith et al. 2001). The presence of this extensive crater profoundly impacts the seasonal dynamics of the region, particularly influencing the recession patterns of the South Polar Seasonal Cap (SPSC) (Colaprete et al. 2005; Giuranna et al. 2007).

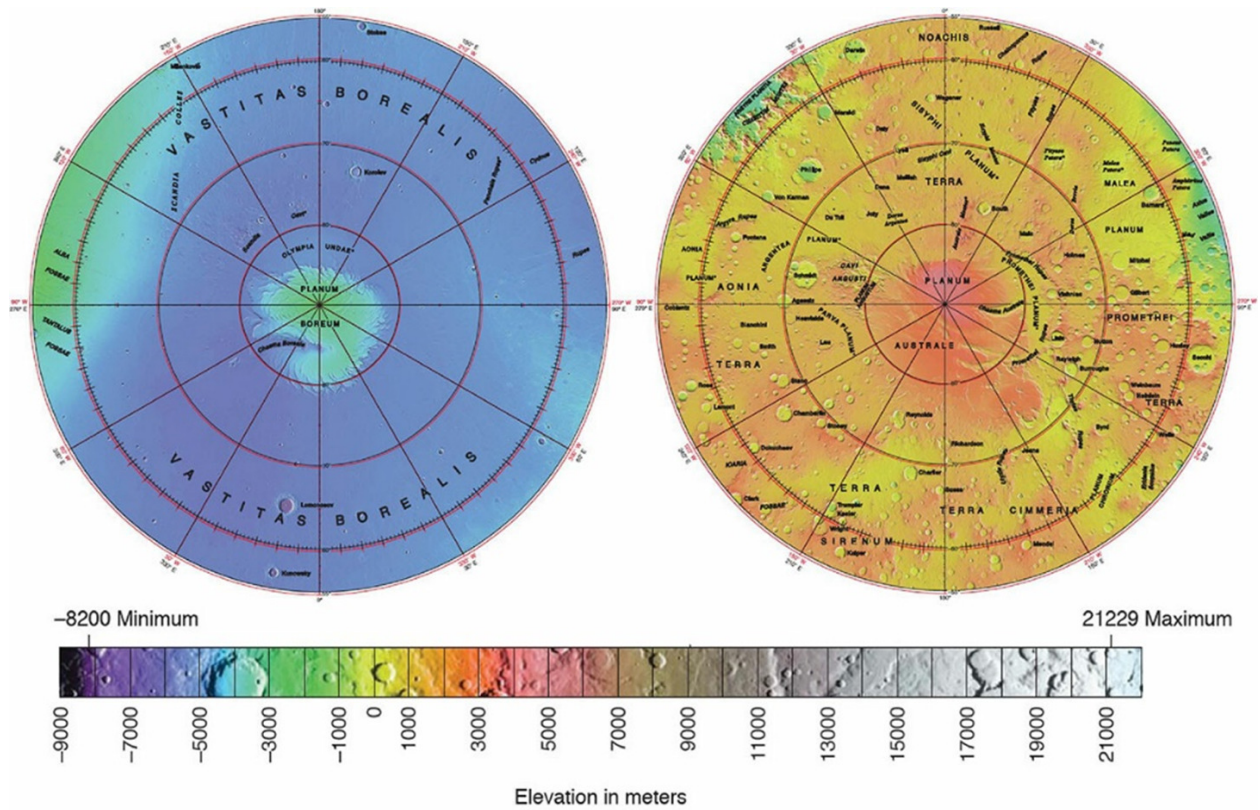


Figure 1.12: Figure 7 from Smith (2022) shows the topographic differences between the northern lowlands and the southern highlands.

1.3.2 Comparison of the North Polar Seasonal Cap and South Polar Seasonal Cap

The North Polar Seasonal Cap (NPSC) and the SPSC display significant differences in their seasonal ice coverage and recession patterns, largely due to Mars' elliptical orbit (Byrne 2009; Haberle 2015). The southern winters are about 24 sols, 24 Martian days, longer than the northern winters (Hansen et al. 2024). The prolonged winters allow for a longer period of CO_2 and water vapour condensation from the atmosphere to form the seasonal polar cap (Hansen et al. 2024). Consequently, the SPSC accumulates a greater amount of ice, in terms of size and mass, making it substantially larger than the NPSC. During maximum extent, the SPSC covers an area of approximately 20 million square kilometers, compared to the NPSC's 17 million square kilometers, marking a significant difference of about 17% (Piqueux et al. 2015b). At the peak extent, the SPSC

contains a peak mass of approximately 6×10^{15} kg compared to the NPSC peak mass of approximately 3×10^{15} kg (Hansen et al. 2024). This size disparity has profound implications for their respective impacts on the Martian climate, including atmospheric pressure variations. Atmospheric pressure changes by more than 30% during the formation and sublimation of the SPSC, in contrast to a variation of about 9% observed with the NPSC (Ordonez-Etxeberria et al. 2018; Martínez et al. 2017) (Figure 1.11).

Another key difference between the SPSC and NPSC is the rate of recession. During the retreat of the SPSC, the surface temperature increases by approximately 3 K per degree of L_s , in contrast to the 2 K per degree of L_s observed during the retreat of the NPSC (Millour et al. 2015). This discrepancy can be attributed to Mars' eccentric orbit, weather models show that the southern polar region receives a total solar irradiance of 280 watts per square meter, whereas the northern polar region receives 255 watts per square meter (Millour et al. 2015). However, the exact amount of total solar irradiance is highly dependent on atmospheric conditions such as optical depth and dust content.

1.3.3 Comparison of the North Polar Residual Cap and South Polar Residual Cap

The North Polar Residual Cap (NPRC) and the South Polar Residual Cap (SPRC) exhibit significant differences in size and reflective properties (Figure 1.6). The NPRC, primarily composed of water-ice, spans approximately one thousand kilometers in diameter and covers most of the North Polar Layered Deposits (NPLD) area (Fishbaugh and Head-III 2001; Smith and Holt 2010; Brown et al. 2016). In contrast, the SPRC is significantly smaller, about five hundred kilometers in diameter, and comprises both CO_2 and water-ice (Fishbaugh and Head-III 2001; Brown et al. 2010). The SPRC is smaller than the NPRC because of the composition, CO_2 has a higher sublimation temperature than water, and the southern summers being warmer than the northern summers. The SPSC is also highly asymmetric and only covers a portion of the South Polar Layered Deposits (SPLD) area (Fishbaugh and Head-III 2001; Giuranna et al. 2007; Smith et al. 2015). These

differences between the residual caps further highlight the distinct polar environments on Mars.

1.3.4 Comparison of the North Polar Layered Deposits and South Polar Layered Deposits

The North Polar Layered Deposits (NPLD) and the SPLD display significant differences in size, composition, and age. Compositionally, the NPLD contain approximately 5% dust, whereas the SPLD have a higher dust content of 10% (Smith 2022). The SPLD are characterized by two distinct CO₂ layers: a thin residual cap on the surface and a more extensive buried layer, adding to their complex structure (Manning et al. 2019; Alwarda and Smith 2021). Furthermore, the physical forms of the layered materials vary between the two polar regions. In the north polar region, the environmental conditions—particularly those related to seasonal and aeolian processes—offer clearer insights into the composition of the NPLD and aid in interpreting Mars' climatic history (Smith 2022). In contrast, the SPLD presents a more complex scenario (Smith 2022). Crater counting studies indicate that the SPLD is between 30-100 million years old, whereas the NPLD is estimated to be about ten thousand years old (Herkenhoff 2000; Koutnik et al. 2002).

1.3.5 Comparison of the North Polar Anomalies and South Polar Anomalies

The understanding of the two polar regions of Mars varies significantly in depth and clarity. The northern polar region is relatively well understood, although both regions present anomalies that have not yet been fully explained in the scientific literature.

In the northern polar region, one such anomaly includes the Cold and Bright Anomalies (CABAs), which manifest as seasonal bright spots on the NPRC during early northern summer, around $L_s = 100^\circ$. Initial investigations by Kieffer and Titus (2001) pinpointed a specific anomaly site at 87° N, 313° E, characterized by a significantly higher albedo and temperatures approximately 10 K colder than the surrounding area. This site, named "Vostok" by Cantor et al. (2010), undergoes considerable changes after $L_s = 147^\circ$, with the albedo decreasing by about 30% and temperatures

rising roughly 5 K above those of the surrounding regions (Kieffer and Titus 2001). Notably, three major darkening events have been documented at this site and the darkened anomaly site consists predominantly of water-ice (Kieffer and Titus 2001; Cantor et al. 2010). The precise mechanisms behind these anomalies remain unclear and are further explored in Chapter Four.

In the southern polar region, the Cryptic Terrain represents a notable anomaly. This anomaly is a portion of the SPSC that has an albedo of approximately 0.25 compared to the higher albedo seasonal cap at approximately 0.45 (Kieffer et al. 2000) (Figures 1.5, 1.13). Spectral studies have confirmed that this region is composed of CO₂ ice (Kieffer et al. 2000; 2006); however, the formation processes of this terrain are not fully understood. This anomaly emerges during the early southern spring (Kieffer et al. 2006; Calvin et al. 2017), and high-resolution surface images reveal fan features with spider-like patterns emanating from the center (Kieffer et al. 2006) (Figure 1.13). It is hypothesized that the anomaly region consists of translucent CO₂ slab ice (Kieffer et al. 2006). The slab ice allows for this region to exhibit a lower albedo and allows for solar radiation to warm up the underlying dark dust layer (Kieffer et al. 2006). The sublimated CO₂ gas is channelled towards a central point, where it breaches the surface ice, creating fan features akin to geysers (Kieffer et al. 2006) (Figure 1.13).

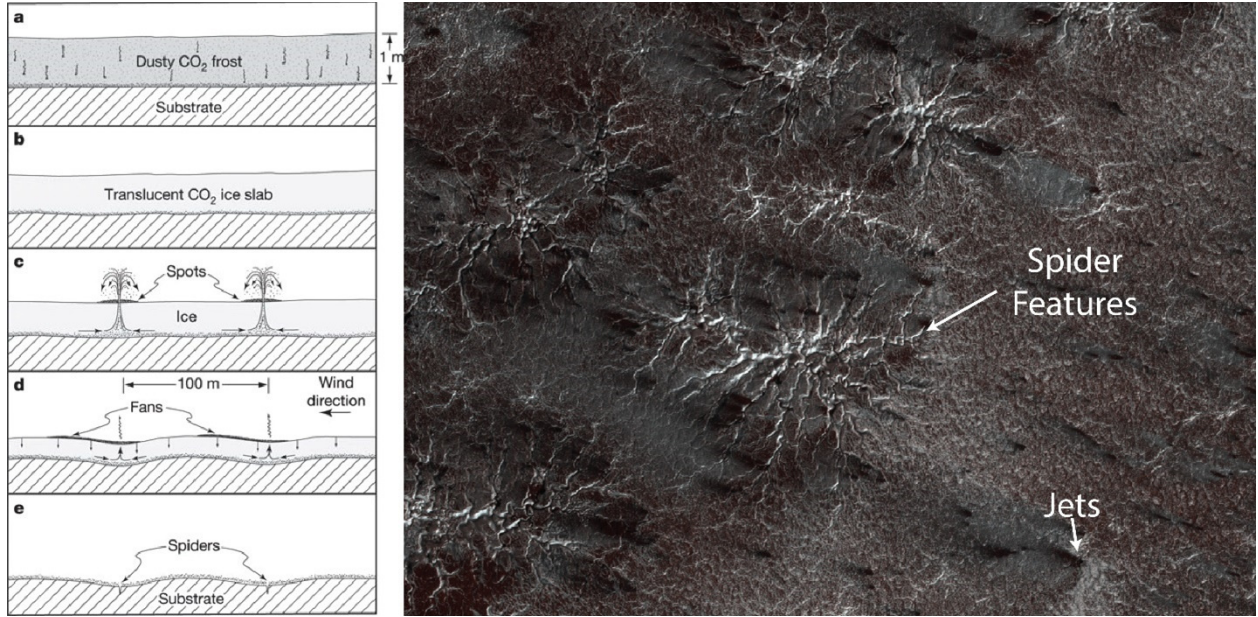


Figure 1.13: Left: Figure 4 from Kieffer et al. (2006) shows a graphical interpretation of the cryptic terrain, fans, and the associated spider features. Right: The High-Resolution Imaging Science Experiment image during MY 29 (2009) at $L_s = 233.4^\circ$ with the jets and spider features marked by arrows. The jets are approximately 500 meters along the long side.

1.4 The Mars Reconnaissance Orbiter

As of 2024, five active Martian orbiters serve as communication satellites for surface missions such as rovers as well as observe the Martian system. These orbiters contain various instruments to observe all aspects of Mars, from magnetometers onboard the MAVEN orbiter to high-resolution imagers with a spatial resolution of less than 10 meters. Out of the seven active orbiters, only three orbiters, the 2001 Mars Odyssey Orbiter, Mars Express, and the Mars Reconnaissance Orbiter (MRO), can observe the polar regions as they are in a solar synchronous, a near-polar orbit that allows a satellite to pass over any given point on the planet's surface at the same solar time. This dissertation heavily relies on observations made from instruments onboard MRO.

1.4.1 An overview on the Mars Reconnaissance Orbiter

MRO is an integral component of NASA's Mars Exploration Program, initiated to explore the presence of water on Mars and to provide comprehensive reconnaissance of the planet's landscape. Launched on August 12, 2005, MRO arrived at Mars on March 10, 2006. Following a series of aerobraking maneuvers, which utilized the Martian atmosphere to decelerate, MRO achieved its stable science orbit, a near-polar trajectory at altitudes between 255 and 320 kilometers, with an orbital period of 112 minutes and an inclination of 86.9° (near-polar) (Zurek et al. 2024). The primary scientific objectives of MRO include:

1. Characterizing Mars' current climate and its seasonal and annual variations (Zurek and Smrekar 2007).
2. Analyzing the global atmosphere and monitoring Martian meteorological conditions (Zurek and Smrekar 2007).
3. Investigating complex Martian terrains to identify water-related geological formations (Zurek and Smrekar 2007).
4. Detecting locations indicative of aqueous or hydrothermal activities (Zurek et al. 2024).
5. Probing the Martian surface for subsurface water and ice and detailing the internal structure of the NPLD and SPLD (Landis et al. 2023).
6. Identifying and characterizing potential sites for future Martian surface explorations (Zurek and Smrekar 2007).
7. Facilitating the relay of scientific data to Earth from surface missions on Mars (Zurek and Smrekar 2007).

MRO is outfitted with a suite of scientific instruments, including three cameras, two spectrometers, and a radar, to fulfill these objectives.

The High Resolution Imaging Science Experiment (HiRISE) camera on MRO, capable of a spatial resolution of 0.3 meters per pixel from an altitude of 300 kilometers, employs a large 160-megapixel imaging chip combined with a 0.5-meter reflecting telescope with a 12-meter focal length, representing the largest telescope dispatched on a deep space mission (McEwen et al. 2007). HiRISE captures images in three colour bands: Blue-Green (400-600 nanometers), Red (550-850 nanometers), and Near-Infrared (800-1000 nanometers). With a swath length of 1.2 kilometers for the Blue-Green and Near-Infrared bands and 6 kilometers for the Red band, HiRISE has been instrumental in monitoring various Martian surface phenomena (McEwen et al. 2007; Landis et al. 2023). Notably, it has observed shallow surface ice exposures caused by recent impact craters smaller than 20 meters in diameter (Landis et al. 2023). Additionally, HiRISE facilitates the generation of Digital Terrain Models (DTMs) through stereo imaging, achieving topographical precision to within 0.25 meters (Landis et al. 2023).

The Context Camera (CTX), another imaging tool aboard MRO, is a grayscale camera that operates between 500 and 800 nanometers, achieving a spatial resolution of about 6 meters per pixel at 300 kilometers, with a swath length of 30 kilometers (Bell et al. 2013; Landis et al. 2023). Comprising a Cassegrain telescope with a 350-meter focal length and a 5064-pixel-wide CCD line array, CTX provides contextual imagery to supplement HiRISE observations and is instrumental in compiling mosaics of extensive Martian regions (Bell et al. 2013; Landis et al. 2023).

In addition, the Mars Colour Imager (MARCI), a wide-angle camera with relatively low resolution, captures Martian surface imagery in both visible and ultraviolet bands. Detailed discussions on MARCI are provided in Section 2.2.1.

The Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) and the Mars Climate Sounder (MCS) are two spectrometers aboard MRO that complement the imaging tools. CRISM, a visual and near-infrared hyperspectrometer, achieves a spectral resolution of 6.5 nanometers and a spatial resolution of 18 meters per pixel at an altitude of 300 kilometers (Murchie et al. 2007; Landis et al. 2023). It has been pivotal in identifying materials indicative of past and present water and hydrothermal activities and in conjunction with HiRISE, aids in producing detailed geological

maps of Mars (Landis et al. 2023). MCS, a radiometer, quantifies the vertical structure of the Martian atmosphere and compiles daily global weather maps, providing vital data on atmospheric conditions, including temperature, pressure, and dust opacity (Landis et al. 2023).

Finally, the Shallow Radar (SHARAD) on MRO, designed to probe up to a kilometer below the Martian surface, emits high-frequency radio waves (15-25 MHz) to map subsurface ice and dust deposits (Landis et al. 2023). SHARAD's observations have been crucial in mapping mid-latitude subsurface ice and in understanding the migration history of spiral troughs within the NPLD and SPLD (Smith et al. 2013; Morgan et al. 2021).

1.5 Background Information on Computer Vision

Computer vision, a field within artificial intelligence, involves enabling computers to interpret and process visual information from the world in the same way humans do (Szeliski 2010). Since the 1980s this technology has evolved significantly, driven by advancements in machine learning, neural networks, and computational power (Szeliski 2010). From its early applications in simple image processing tasks in the 1960s, computer vision now plays a critical role in various industries, including healthcare, automotive, and surveillance (Szeliski 2010; Wagstaff et al. 2018). The ability of computer vision to analyze vast amounts of visual data quickly and makes it indispensable in modern times.

The importance of automation in space missions cannot be understated, especially when considering challenges posed by remote operations and communication delays between Earth and Mars. Computer vision technology allows for autonomous navigation of rovers, enabling them to traverse the Martian terrain efficiently. For example, the Martian Exploration Rover mission, which deployed the Spirit and Opportunity rovers, integrated advanced computer vision algorithms for autonomous navigation and hazard avoidance (Matthies et al. 2007). By utilizing stereo images, these rovers can perceive the surroundings in three dimensions, allowing them to avoid unseen obstacles and continue to explore the surface without waiting for instructions from Earth (Matthies

et al. 2007).

Scientific applications of computer vision often extend to feature detection and classification, which may combine computer vision algorithms with machine learning techniques. Fanara et al. (2020) utilized computer vision and machine learning to autonomously detect block falls at the edge of the NPLD. In that study, computer vision is used to extract various surface features such as crater rims and cliff edge of the NPLD. This information is then used by a neural network to determine which image contains the block falls of interest. The autonomous method is significantly faster than manually analyzing high-resolution images and a small percentage of images are required to be manually inspected due to false positives and false negatives. Similarly, Mishev and Smith (2022) utilized computer vision and machine learning to detect and map surface features associated with bright deposits found near the Valles Marineris, located at the equator. Mishev and Smith (2022) used computer vision to extract features associated with the bright deposits such as shape, colour and general characteristics, which were then classified by a neural network. Similar to Fanara et al. (2020), the automated detection of features associated with bright deposits is significantly faster than manual methods and reduces the number of images that need to be manually inspected. Additionally, the maps generated by the automated method provide key insights into the formation of these features and the history of surrounding areas.

1.6 Dissertation Objectives

This dissertation aims to enhance the understanding of Martian polar regions through three objectives. Each objective utilizes orbital observations that may be combined with weather simulations, to examine a specific aspect of the Martian polar regions.

Objective #1: Seasonal Variability of the North Polar Seasonal Cap:

The first objective of the dissertation is to analyze the retreat of the SPSC during the northern spring, $L_s = 0^\circ$ to $L_s = 70^\circ$ for multiple MYs. The NPSC is one of the largest surface features and the most influential on Mars, capable of modulating the global atmospheric pressure. Typically, the

analysis of the recession of the NPSC relies on manual method, especially for visual band images, which can take several weeks to complete for images spanning multiple MYs (Calvin 2022).

To achieve this, a novel computer vision algorithm is used to track the NPSC using daily observations from MYs 29-35. These observations are critical for determining the annual recession rates and identifying new sub-seasonal trends previously obscured by limitations in manual processing and the number of observations used.

The impact of this objective is multifaceted:

1. **Enhanced Efficiency and Accuracy:** By automating the tracking process, the novel method aims to significantly reduce the time required for analysis, increasing the efficiency and accuracy of seasonal recession rates.
2. **Improved Climate Model:** The high-resolution data obtained will refine current climate models of Mars, offering more precise predictions of future climatic conditions.
3. **Sub-Seasonal Trends Identification:** The ability to identify new sub-seasonal trends provides deeper insights into the dynamic processes associated with the NPSC.
4. **Broader Implications for Martian Science:** The findings associated with this objective will enhance our understanding of the Martian polar processes and how the north polar region can impact the broader Martian climate, particularly in the study of seasonal variability of mid-latitude regions.

Objective #2: Seasonal Variability of the South Polar Seasonal Cap:

The second objective of this dissertation is to analyze the retreat of the SPSC during the southern spring, $L_s = 180^\circ$ to $L_s = 270^\circ$. The SPSC is 17% larger than the NPSC and has been observed to have a larger impact on the global atmospheric pressure than the NPSC. Previous works monitoring the retreat of the SPSC utilized manual outlines that are computationally slow and many observations were ignored to reduce the processing time.

To achieve this, a novel computer vision algorithm is used to track the retreat of the SPSC using daily observations from MYs 28-31. These observations are essential for determining the annual

recession rate and identifying the variability in the retreat of the SPSC. The impact of this work is very similar to the impact of the first objective.

Objective #2: Hypothesizing the formation of the NPRC Anomalies:

The third objective of this dissertation is to develop a comprehensive hypothesis for the CABAs observed on the NPRC during polar summer. This anomaly has been observed for several decades; however, the formation is unknown (Kieffer and Titus 2001; Cantor et al. 2010). This comprehensive hypothesis will be the first all-encompassing explanation for the formation and evolution of the CABAs.

The impact of this research is significant:

1. **New Insights into Martian Polar Processes:** Developing a thorough understanding of the CABAs will offer valuable insights into the poorly understood underlying physical and climatic processes occurring in the Martian polar regions.
2. **Enhancing Predictive Models:** A detailed hypothesis will provide new polar processes that are not fully captured in current weather models. This can impact future Martian missions, specifically surface mission, as weather models play a crucial role in the planning stage of a successful mission.

Chapter 2

Tracking the North Polar Seasonal Cap Retreat of Mars Using Computer Vision

2.1 Motivation

The Northern Polar Seasonal Cap (NPSC) is among the most prominent surface features on Mars, observable through both telescopes and orbital imagers. At its peak during the northern winter solstice, approximately at $L_s = 270^\circ$, the NPSC spans an area of about 17 million square kilometers, extending down to a latitude of 51° N (Piqueux et al. 2015b). This vast ice cap plays a crucial role in modulating the global atmospheric pressure and composition, as it accounts for the condensation of up to 25% of the Martian atmosphere during the northern winter, which subsequently sublimates in the northern spring (Tillman et al. 1993). Additionally, the periphery of the NPSC is a frequent site for regional dust storms, which can extend thousands of kilometers, sometimes reaching as far as the equator. These storms often co-occur with smaller water-ice clouds, further influencing Martian climatic dynamics (Cantor et al. 2010; Calvin et al. 2015).

The primary objective of this chapter is to enhance the methodologies previously established by Calvin et al. (2015). Calvin et al. (2015) utilized mosaicked observations from the Mars Color Imager (MARCI) to analyze the north pole, spanning Solar Longitude (L_s) = 0° to $L_s = 80^\circ$ (northern

spring) for Mars Years (MY) 29-31. This analysis primarily aimed to determine the extent of the NPSC. To reduce the processing time, Calvin et al. (2015) limited their analysis for every 5° of L_s (Solar Longitude) dates, covering about 20% of the observations for a given MY. Furthermore, they employed a circle of best fit to characterize the NPSC extent, a method that becomes less accurate as the NPSC begins to retreat asymmetrically in late spring

This chapter aims to develop an autonomous method that utilizes the same MARCI mosaics used by Calvin et al. (2015). This new approach aims to significantly speed up the analysis process compared to manual methods and will incorporate nearly the entire MARCI dataset, thereby enhancing the temporal resolution of the seasonal recession curve. By utilizing more observations, this method may reveal unresolved sub-seasonal trends, offering a deeper insight into the seasonal dynamics of the NPSC retreat. Additionally, this chapter will extend Calvin et al. (2015) work by including MARCI mosaics from MYs 32-35, aiming to identify seasonal reoccurring sub-seasonal trends and possibly new seasonal trends in the new MYs.

2.2 Methods

2.2.1 MARCI Dataset

This chapter uses polar mosaicked images from the Mars Colour Imager (MARCI) from MYs 29 to MY 35, spanning $L_s = 0^\circ$ to $L_s = 170^\circ$, which were constructed by Calvin et al. (2015).

The Mars Color Imager (MARCI), onboard the MRO, is a wide-angle (180° field-of-view) push-frame camera operational since MY 28 (2006) $L_s = 111.4^\circ$ (Malin et al. 2001). The primary objective of the MARCI camera is to aid climatological and meteorological studies by monitoring the atmosphere and polar processes (Malin et al. 2001). MARCI is an enhancement of the earlier Mars Orbiter Camera onboard the Mars Global Surveyor launched in 2001. Mars Orbiter Camera has an average resolution of 3.8 kilometers per pixel, compared to MARCI's average resolution of 1.9 kilometers per pixel (Cantor et al. 2010; Malin et al. 2001). One of MARCI's primary functions is to provide insights into the surface and atmospheric conditions (Malin et al. 2001).

MARCI is equipped with seven colour bands ranging from 437 nanometers to 718 nanometers and two UV bands at 258 nanometers and 320 nanometers (Bell et al. 2009). This instrument captures images with a resolution of approximately 930 meters per pixel from an altitude of 300 kilometers (Malin et al. 2001; Bell et al. 2009). However, due to bandwidth limitations and storage space, the images were downsampled to a resolution of 1.9 kilometers per pixel. The camera systematically acquires images from pole to pole along the sunlit side of Mars at approximately 15:00 local mean solar time, covering the entire ground track from limb to limb (Calvin et al. 2015). Overlap between adjacent orbits is minimal at the equator but substantial at the poles, facilitating enhanced imaging of these regions (Cantor et al. 2010).

The camera completes about 12–13 orbits daily, with a significant focus on the polar regions, enabling synoptic coverage that captures the dynamic changes in polar albedo deposits (Cantor et al. 2010). MARCI data are processed into daily global maps at a resolution of 1.9 kilometers per pixel, applying Minnaert photometric correction and projecting into polar stereographic coordinates (Calvin et al. 2015). Data from filter band 3 (604 nanometers), band 2 (546 nanometers), and band 1 (437 nanometers) were utilized to synthesize an RGB image (Cantor et al. 2010; Calvin et al. 2015).

During image processing, frames with significant artifacts or insufficient polar coverage are excluded (Calvin et al. 2015). The remaining images are mosaicked into daily global maps, which are further utilized to create time-sequence movies showcasing seasonal changes over multiple MYs (Calvin et al. 2015). The images are time-stamped in Earth days, with the L_s value for each mosaic set at the midpoint of the day to accommodate slight variations during image acquisition (Calvin et al. 2015).

For determining the extent of the Northern Polar Seasonal Cap (NPSC), we convert the pixel values from the MARCI mosaics into latitude degrees. This conversion is crucial due to the modified product is slightly larger than the native scale of 1.9 kilometers per pixel. Employing a polar stereographic projection, where lines of longitude radiate outward and rings of latitude form concentric circles around the geographic North Pole, we establish that the latitude offset from the

center, representing the pole, remains consistent across the mosaic. This uniformity allows us to adopt a constant conversion factor from pixels to latitude degrees.

To accurately establish this factor, we utilize the JMARS software to measure the latitude distances from the geographic pole to six identifiable surface features, such as craters (Figure 2.1). JMARS (Java Mission-planning and Analysis for Remote Sensing) is a mission planning software that allows users to visualize observations, display maps, and other data analysis tools (Christensen et al. 2009). We determine the pixel distances to these same features using the polar mosaic. By calculating the ratio of latitude degrees to pixels for each site, we derive an average conversion ratio of 0.061° per pixel or about 3.6 kilometers per pixel. This ratio will be consistently applied throughout this chapter to convert all mosaic measurements into geographical latitude, ensuring precise tracking of the NPSC's coverage.

2.2.2 Outlining the North Polar Seasonal Cap

The entire automation is developed in Python and uses an open-source Python library called OpenCV for the majority of the image processing and feature detection (Bradski 2000).

We first convert the Red, Green, and Blue (RGB) MARCI images to Hue, Saturation, and Value (HSV) colour format using the `cvtColor` algorithm found in OpenCV. In this colour format, low H values represent red hues, while high H values represent blue hues (Figure 2.1). Low S values indicate low saturation, while high S values denote highly saturated colours (Figure 2.1). Similarly, low V values represent dark grayscale values, and high V values denote bright grayscale values (Figure 2.1).

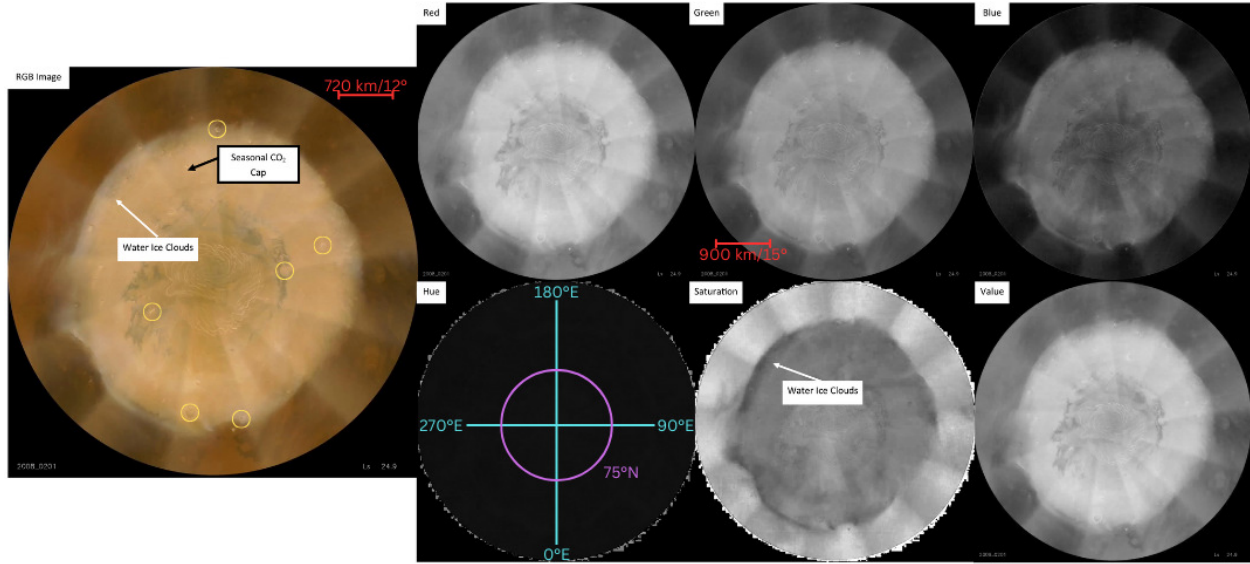


Figure 2.1: MARCI images at MY 29 $L_s = 24.9^\circ$ broken into various characteristics Left: Full-color MARCI mosaic with the yellow circles marking the craters used to determine the pixel conversion rate Top row: R, G, and B channels. There is an insufficient distinction between the three channels to determine where the water and CO_2 ice boundaries are. Bottom row: the H, S, and V channels with the center of the image marking 90°N and purple circle marking the 75°N , roughly the extent of the NPDL. The Hue (mostly red) is very low, and V is unable to distinguish between CO_2 and water-ice boundaries. In the S channel, water-ice clouds are darker than the seasonal cap, providing a boundary that can be automatically detected.

Examining the individual channels closely (Figure 2.1), the H channel shows low values, indicating that the northern region is dominated by red hues. The S channel reveals that the unfrosted region is very saturated, the seasonal cap is less saturated, and water-ice clouds between the seasonal cap and the defrosted regions are the least saturated. The V channel displays high values for all types of ice but not the surrounding plains. Using the S channel, by setting appropriate upper and lower bounds, we select for the regions that neither exhibit the water-ice clouds nor the surrounding plains. A benefit of HSV decomposition is that the V or brightness, channel is separate, making our detections robust to external light changes—such as the entire region brightening as the solar zenith angle decreases. To identify the NPSC, we use the `inRange` algorithm from OpenCV. This algorithm requires lower and upper bounds of the H, S, and V channels, which we empirically determined to be (14, 58, 166) and (72, 163, 255), respectively. The algorithm outputs a binary image, where the 255 value represents pixels that satisfy the HSV bounds and 0 where it fails (Figure 2.2a).

Next, we extract contours by finding where the binary image (of points detected with the `inRange` algorithm) transitions from 0 to 255. We use the `findContour` algorithm in OpenCV with the `CHAIN_APPROX_SIMPLE` input parameter. The algorithm draws a straight line from an arbitrary starting point to a nearby contour point such that the white regions in the binary images are not clipped, not cutting into the white region. Then, it removes all contour points present within the line; this process is repeated until it returns to the starting point. This algorithm generates numerous contours (Figure 2.2b). To isolate the largest area contour, the one that represents the NPSC, we first find the area of the individual contours, in pixels, using OpenCV's `contourArea` algorithm and then select the contour with the largest area value (Figure 2.2c). This eliminates smaller, internal contours that surround small features or image artifacts. We did try other methods to isolate this contour but found that the largest area contour is the most consistent at outlining the seasonal cap.

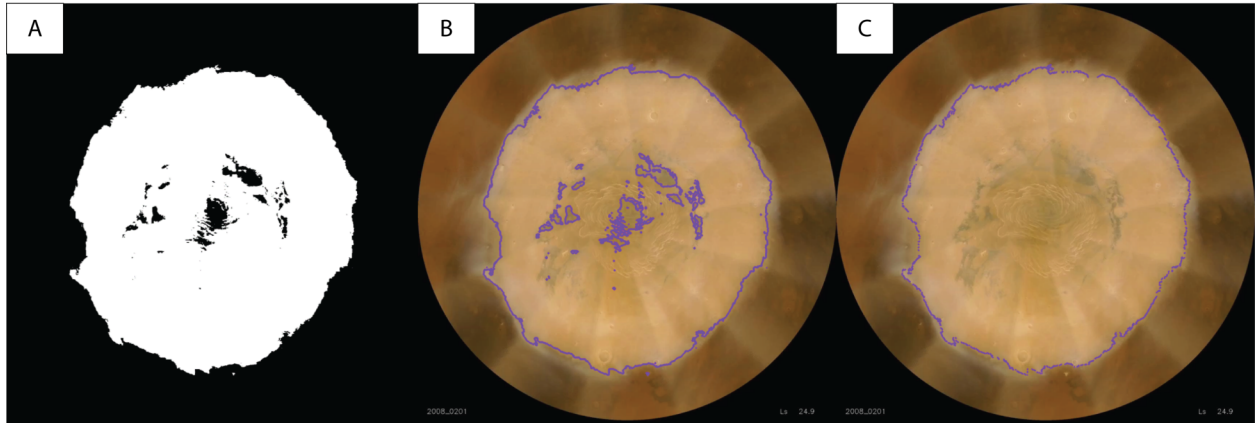


Figure 2.2: Output images from the automated method starting with Figure 2.1. A: The binary image output where white represents pixels that satisfy our empirically derived H, S, and V bounds. B: The contours found by the `findContour` algorithm. C: The largest area contour that best outlines the seasonal cap.

2.2.3 Ellipse Fitting

We employ the OpenCV `fitEllipse` algorithm to find the best fitting ellipse to the cap edge contour. As mentioned in Section 2.1, the circle of best fit is not accurate at characterizing the extent of the NPSC. The ellipse fits the next simplest shape with a higher degree of freedom, which allows it to characterize the asymmetric retreat of the NPSC. The difference between the circle of best fit and the ellipse fit is explored in more detail in Section 2.2.5.

This algorithm uses a least-squares method to determine the center of the ellipse, the major and minor axis, and the azimuth from 0° longitude for a given outline and then draws the ellipse (Figure 2.3a). We chose an ellipse of best fit rather than a circle because an ellipse offers more degrees of freedom, allowing it to better conform to the non-circular shape of the cap. This is particularly true during late spring when the cap retreats asymmetrically.

Several MARCI mosaics contain image processing artifacts, anomalous albedo evolution of the seasonal cap, or regional dust storms that alter the H, S, and V values of the seasonal cap outside the selected range. This affects the ellipse fit (red outlines in Figures 2.3b-d). To address this, we develop an iterative method to filter out any bad contour points by comparing the distances of the contour points to the center of the ellipse. Points that are more than 80 pixels from the original ellipse are excluded in the next iteration of fitting (red points in Figures 2.3b-d). This filtering significantly enhances the robustness of the ellipse fits and allows me to retain some mosaics that would otherwise be excluded because they returned faulty ellipses. To determine the average axis latitude, we then average the major and minor ellipse axes (Eq 2.1) and convert them to latitude.

$$R_{\text{ellipse}} = \frac{2a + b}{3} \quad (2.1)$$

a is the semi-major axis and b is the semi-minor axis of the fitted ellipse fit. In this analysis, we employ a weighted average axis of the ellipse instead of the conventional average axis, $(a+b)/2$, as it more accurately captures the inherent asymmetry of the ellipse—a critical factor for this method. The weighted average axis is geometrically significant, as it incorporates the distinct contributions of both axes in defining the ellipse’s characteristics, rather than merely approximating it as near-circular. The formula for the weighted average axis emphasizes the semi-major axis, which represents the maximum extent of the NPSC, while also acknowledging the influence of the semi-minor axis. This approach ensures a more accurate representation of the elliptical properties, critical for detailed analysis and modelling of the NPSC’s spatial dynamics.

We monitor how the area of the seasonal cap changes with L_s by finding the area of the fitted

ellipse using the area of the ellipse (Eq 2.2) and the arclength of a circle (Eq 2.3).

$$A = \pi ab \quad (2.2)$$

Where A is the area of the ellipse, a and b are the semi-major and semi-minor axes of the ellipse in kilometers. To find the latitude extent of the cap, we record the average axis of the ellipse fit using the following equation.

$$s = 2\pi r \left(\frac{\theta}{360} \right) \quad (2.3)$$

Where s is the arclength in kilometers, r is the polar radius of Mars 3373.3 kilometers (Haberle, 2015), and θ is the latitude displacement between a point on the ellipse and the geographic north pole.

We repeat these steps every day until $L_s = 70^\circ$, around the time when the NPSC reaches the NPLD, and isolating the seasonal cap from the polar ice becomes difficult. Using a 2021 MacBook Air with the Apple M1 8-core CPU, we process seven seasons' worth of MARCI observations in about 2.5 minutes, and each MY takes 15 seconds. Calvin et al. (2015) hand-selection approach required several weeks of personnel time to outline a limited subset of the MY 29 to 31 images. This speed allows me to rapidly adjust the code or bounding values for HSV to improve the empirical fit and then analyze all of the MARCI images very quickly. The result is high temporal fidelity and an ability to resolve sub-seasonal trends.

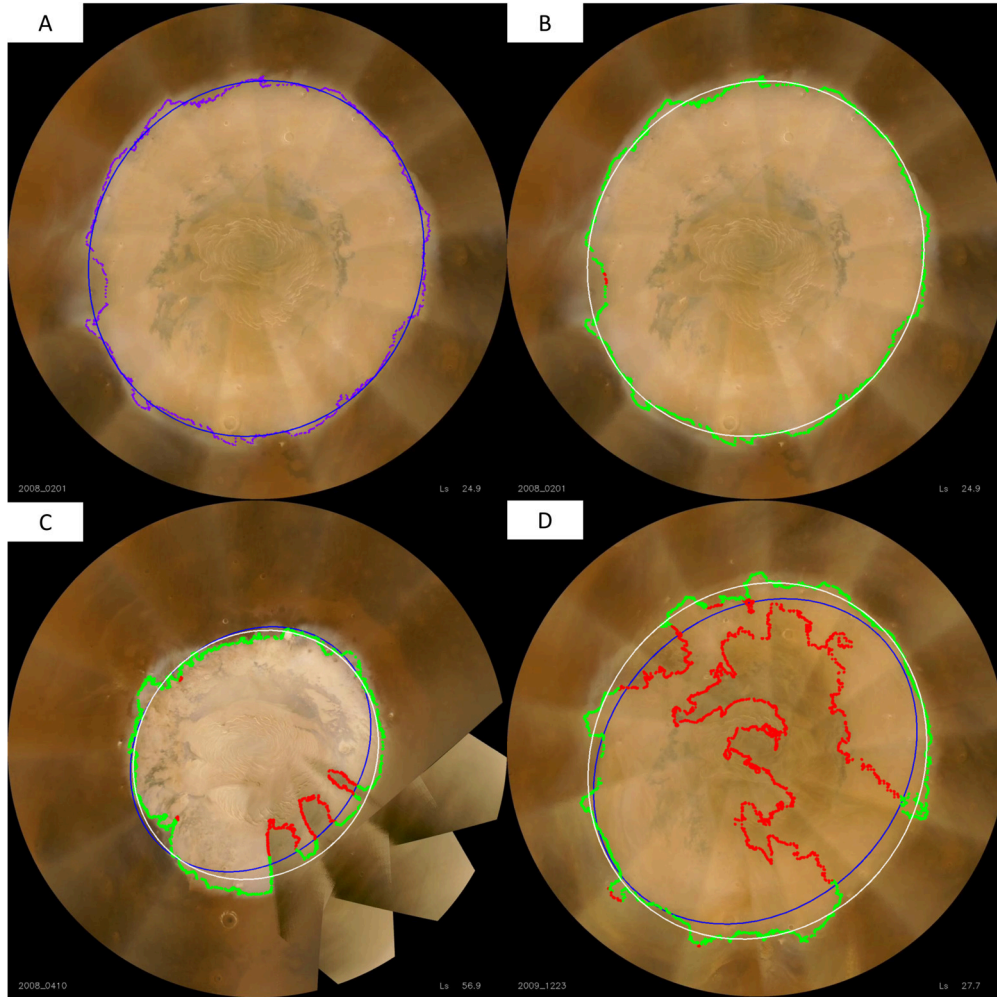


Figure 2.3: MARCI mosaic (from Figure 2.2a) with an initial ellipse fit (Blue) considering all the contour points (Purple). B: The same MARCI mosaic with the filtered ellipse fit (White) using the points that passed the filtering algorithm (Green) and ignoring the points that failed (Red). C: A MARCI mosaic, contour, and two ellipses for MY 29 $L_s = 56.9^\circ$, with data processing artifacts. An initial ellipse fit (blue) and filtered ellipse fit (White) removed red points that deviate strongly from the original ellipse. D: MARCI mosaic during a regional dust storm at MY 30 $L_s = 27.7^\circ$. Red points that go into making the blue ellipse are filtered out in forming the white ellipse.

2.2.4 Dealing with Outliers

There were a few mosaics in which the filtered ellipse fit did not accurately outline the seasonal cap. This is largely due to regionally intensive dust activity, which obscured the cap and altered its H, S, and V values. There were also a few outliers resulting from data gaps, where parts of the seasonal cap were missing, or due to distortions from artifacts of the image processing. Most of

the outliers associated with regional dust storms occurred in MY 32 from $L_s = 23^\circ$ to $L_s = 24^\circ$. We removed two mosaics during early spring, $L_s = 26.8^\circ$ and $L_s = 27.3^\circ$ from the MY 32 dataset, which represents approximately 1.5% of the data from that year. All plots in this chapter have these outliers removed.

2.2.5 Validation

We validate our method by comparing our results with those from (Calvin et al. 2015), examining the outlines of the cap (Figure 2.4a) and the recession rates for MY 29, 30, and 31 (Figures 2.4b-d; Table 2.1). The differences are minor, and the largest discrepancy occurs for MY 29, where our average linear fit and that of Cantor et al. (2010) are approximately 2.5° less advanced than that of Calvin et al. (2015) 2.4b). Cantor et al. (2010) had manually determined the extent of NPSC for MYs 28-29. Linear fits for both MY 30 and MY 31 align with previous works (Figures 2.4c-d).

Certain areas exhibit significant discrepancies between the circle of best fit and the actual outline of the NPSC, notably the region identified as number 1 in Figure 2.4a, where the difference amounts to 0.22° N. The limitations of the circle of best fit arise from its reliance on only two parameters: the center point and the radius, which restricts its ability to accurately represent the asymmetrical shape of the NPSC. Conversely, the ellipse of best fit provides a more accurate characterization due to its incorporation of four parameters: the center point, the semi-major axis, the semi-minor axis, and the rotation angle. This additional complexity allows for a finer adjustment to the actual contours of the NPSC. In the same region marked by number 1 in Figure 2.4a, the difference using the ellipse of best fit is reduced to 0.098° N, representing a 44% improvement over the circular fit. Despite this advancement, the ellipse of best fit still does not perfectly encapsulate the NPSC's asymmetry. Its structural limitations restrict its ability to fully conform to the cap's irregular contours.

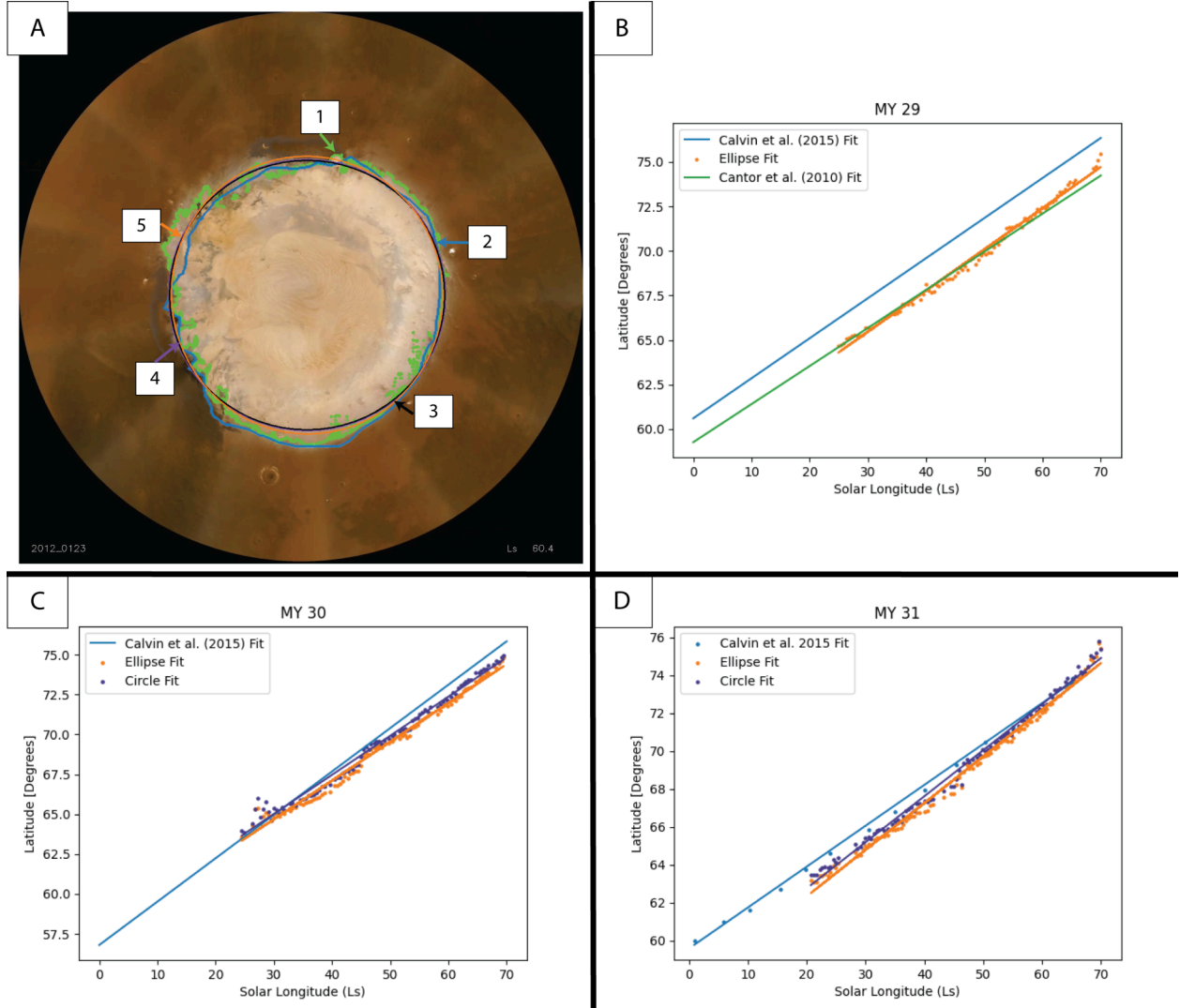


Figure 2.4: Comparison between our results and those from Cantor et al. (2010) and Calvin et al. (2015) for a MARCI mosaic at MY 29 $L_s = 62.5^\circ$. A: (1) our outline in green, (2) Calvin et al. (2015)'s outline in blue, (3) Calvin et al. (2015)'s average latitude in black, (4) our circle of best fit using the HSV method in purple, and (5) our ellipse of best fit using the HSV method in orange. B: Comparisons of the three survey's latitudinal extents over MY 29. Calvin et al. (2015) in Blue; our best fit ellipse (Orange); our circle of best latitude fit (Purple); and Cantor et al. (2010) in Green. C and D: Same as B but for MYs 30 and 31.

Table 2.1: Comparison of the lines of best with the automated method and the manual methods used by Calvin et al. (2015), and Cantor et al. (2010)

MY	Our Fit	Calvin et al. (2015) fit	Cantor et al. (2010) best-fit
29	$(0.22 \pm 0.002)L_s + (59.63^\circ \pm 0.10^\circ)$	$0.23L_s + 60.6^\circ$	$0.21L_s + 59.26^\circ$
30	$(0.24 \pm 0.003)L_s + (58.49^\circ \pm 0.14^\circ)$	$0.27L_s + 56.8^\circ$	—
31	$(0.23 \pm [0.002]L_s + (58.51^\circ \pm 0.11^\circ)$	$0.22L_s + 60.0^\circ$	—

2.2.6 Fits

To analyze the output from our algorithms, we fit the data in several ways, including plots of area of the NPSC versus L_s and latitude of the NPSC versus L_s (Figures 2.5a-b). To enhance any trends, we utilize a line of best fit, a moving average, a 7th-order polynomial fit, and its first derivative. Full equations for the linear and the 7th-order fit for the area and latitude recession can be found in the Appendix in Tables 6.1 & 6.2. The line of best fit most closely approximates previous results and provides general characteristics of the recession (Tables 2.2 & 2.3). For the moving average, we use a group size of 5 points or 2.5° of L_s dates, which highlights short-period deviations from the linear fit while reducing noise. we find that the 7th-order polynomial best characterizes all the small deviations occurring during the observational period. We also examine the first derivative of the 7th-order fit to characterize how the recession rate changes with L_s , providing another way to monitor the recession of the cap (Figures 2.5c-d).

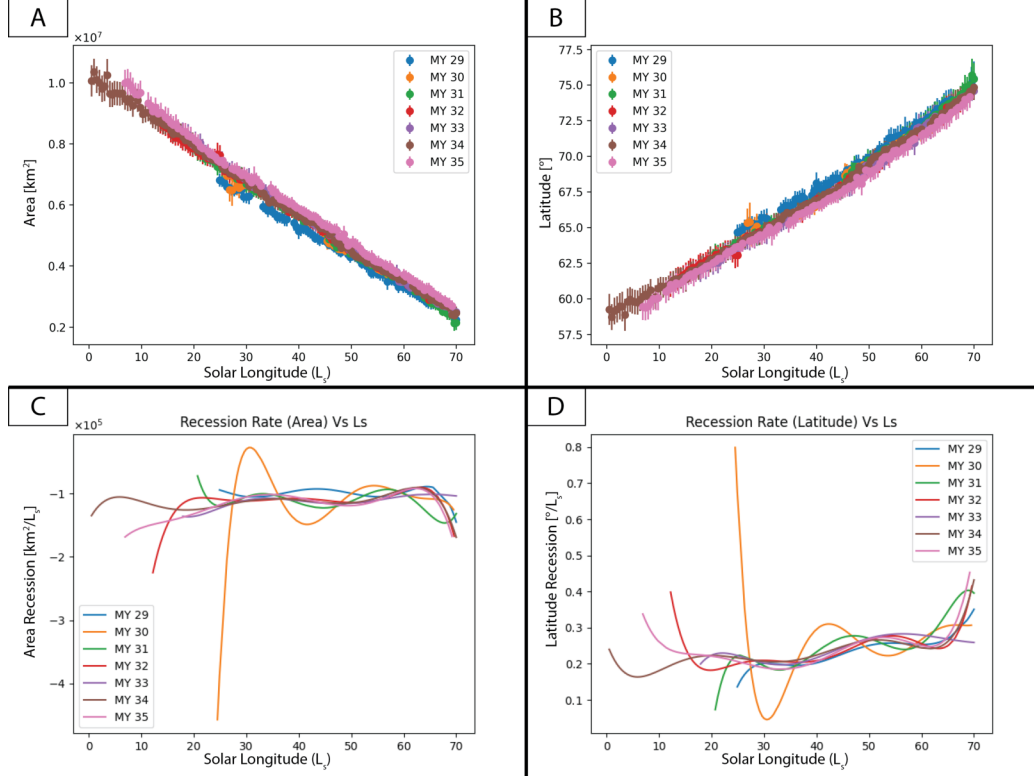


Figure 2.5: The combined seasonal variability of the NPSC area, latitude, and the respective recession rate for MYs 29-35 from $L_s = 0^\circ$ to $L_s = 70^\circ$. A: NPSC Area (Square Kilometers) vs L_s . B: NPSC average axis (Latitude °) vs L_s . C: NPSC area recession rate (Square Kilometer per L_s) vs L_s . D: NPSC latitude recession rate (Latitude Degree per L_s) vs L_s .

2.3 Results

This section outlines the average and individual annual results from the seven MYs, showing the behaviour of the average axis and the area of the fitted ellipse versus L_s (Figure 2.6; Tables 2.2 & 2.3), along with their derivatives. Generally, each year followed very similar trends with comparable linear recession rates; however, we have observed several sub-seasonal and inter-annual variations.

Table 2.2: Table showing the NPSC latitude line of best fit and important dates. Min. latitude recession column shows the date for the annual minimum recession rate and the rate for each column. Max. latitude recession column shows the date for the annual maximum recession rate and the rate for each column

MY	Line of best fit	Min. latitude recession		Max. latitude recession	
		Date	Rate ($^{\circ}/L_s$)	Date	Rate ($^{\circ}/L_s$)
7-year average	$(0.23 \pm 0.0019)L_s + (57.88^{\circ} \pm 0.102^{\circ})$	$L_s = 1.8^{\circ}$	0.12	$L_s = 51.6^{\circ}$	0.27
29	$(0.23 \pm 0.0021)L_s + (58.53 \pm 0.105^{\circ})$	$L_s = 36.8^{\circ}$	0.20	$L_s = 53.1^{\circ}$	0.26
30	$(0.24 \pm 0.0030)L_s + (57.38 \pm 0.147^{\circ})$	$L_s = 30.0^{\circ}$	0.05	$L_s = 42.0^{\circ}$	0.31
31	$(0.25 \pm 0.0023)L_s + (57.40 \pm 0.112^{\circ})$	$L_s = 32.8^{\circ}$	0.18	$L_s = 69.0^{\circ}$	0.40
32	$(0.23 \pm 0.0016)L_s + (58.02 \pm 0.071^{\circ})$	$L_s = 19.5^{\circ}$	0.18	$L_s = 54.1^{\circ}$	0.28
33	$(0.24 \pm 0.0018)L_s + (57.41 \pm 0.085^{\circ})$	$L_s = 33.1^{\circ}$	0.20	$L_s = 56.2^{\circ}$	0.29
34	$(0.23 \pm 0.0013)L_s + (58.24 \pm 0.053^{\circ})$	$L_s = 5.7^{\circ}$	0.16	$L_s = 51.4^{\circ}$	0.27
35	$(0.23 \pm 0.0015)L_s + (57.77 \pm 0.062^{\circ})$	$L_s = 33.2^{\circ}$	0.19	$L_s = 52.7^{\circ}$	0.27

Table 2.3: Table showing the NPSC area line of best fit and important dates. Min. latitude recession column shows the date for the annual minimum recession rate and the rate for each column. Max. latitude recession column shows the date for the annual maximum recession rate and the rate for each column

MY	Area linear slope (km^2/L_s)	Min. area recession		Max. area recession	
		Date	Rate (km^2/L_s)	Date	Rate (km^2/L_s)
7-year average	$(-1.12 \times 10^5 \pm 575)$	$L_s = 16^{\circ}$	-1.34×10^5	$L_s = 61.8^{\circ}$	-9.21×10^4
29	$(-9.89 \times 10^4 \pm 492)$	$L_s = 30.6^{\circ} \& 55.0^{\circ}$	-1.05×10^5	$L_s = 64.9^{\circ}$	-8.89×10^4
30	$(-1.07 \times 10^5 \pm 1127)$	$L_s = 41.1^{\circ}$	-1.48×10^5	$L_s = 30.4^{\circ}$	-2.49×10^4
31	$(-1.11 \times 10^5 \pm 518)$	$L_s = 67.8^{\circ}$	-1.47×10^5	$L_s = 57.0^{\circ}$	-9.33×10^4
32	$(-1.01 \times 10^5 \pm 435)$	$L_s = 51.0^{\circ}$	-1.15×10^5	$L_s = 63.2^{\circ}$	-9.00×10^4
33	$(-1.12 \times 10^5 \pm 448)$	$L_s = 18.8^{\circ}$	-1.37×10^5	$L_s = 65.7^{\circ}$	-1.01×10^5
34	$(-1.13 \times 10^5 \pm 419)$	$L_s = 18.3^{\circ}$	-1.26×10^5	$L_s = 65.7^{\circ}$	-9.04×10^4
35	$(-1.14 \times 10^5 \pm 586)$	$L_s = 49.5^{\circ}$ ⁴³	-1.96×10^5	$L_s = 62.4^{\circ}$	-9.19×10^4

2.3.1 Repeatable Trends, Seven-year Average, and Annual Results

The linear fit for each year serves as a rough approximation of the recession rates; however, we use it as a basis to guide discussions for comparing the cap to an annual trend. Generally, the recession rate (measured by average latitude) is slower earlier in the season and faster later in the season. This creates acceleration in the latitude points (Figures 2.6a-c) or an upward slope in the recession rate (Figure 2.6b), first identified by Kieffer and Titus (2001).

We notice that before $L_s = 26.7^\circ$, the latitudinal extent is always smaller (higher latitude) than the linear fit implies (Figure 2.6a). Likewise, later in the season, starting around $L_s = 45^\circ$, but always by $L_s = 60^\circ$, and continuing until the last data point, the cap is also smaller than the linear fit implies. The middle part of the season (black arrows in Figure 2.6a) balances the ends by showing a larger cap than the linear fit implies. Looking at the latitude recession rate (Figure 2.6b), there is a relatively slow period at the beginning of the year that shifts to a relatively fast period starting between $L_s = 45^\circ$ and $L_s = 60^\circ$ and continuing to the end. This acceleration is most obvious in MY 30 (Figures 2.7a,b).

Regarding the area plots, the same trends are identifiable, although with opposite descriptors because as the cap's latitude increases, the area decreases. The area of the cap tends to be smaller than the linear fit until $L_s = 22^\circ$ and again towards $L_s = 60^\circ$. In the middle of the season, the NPSC is to be larger than the linear fit.

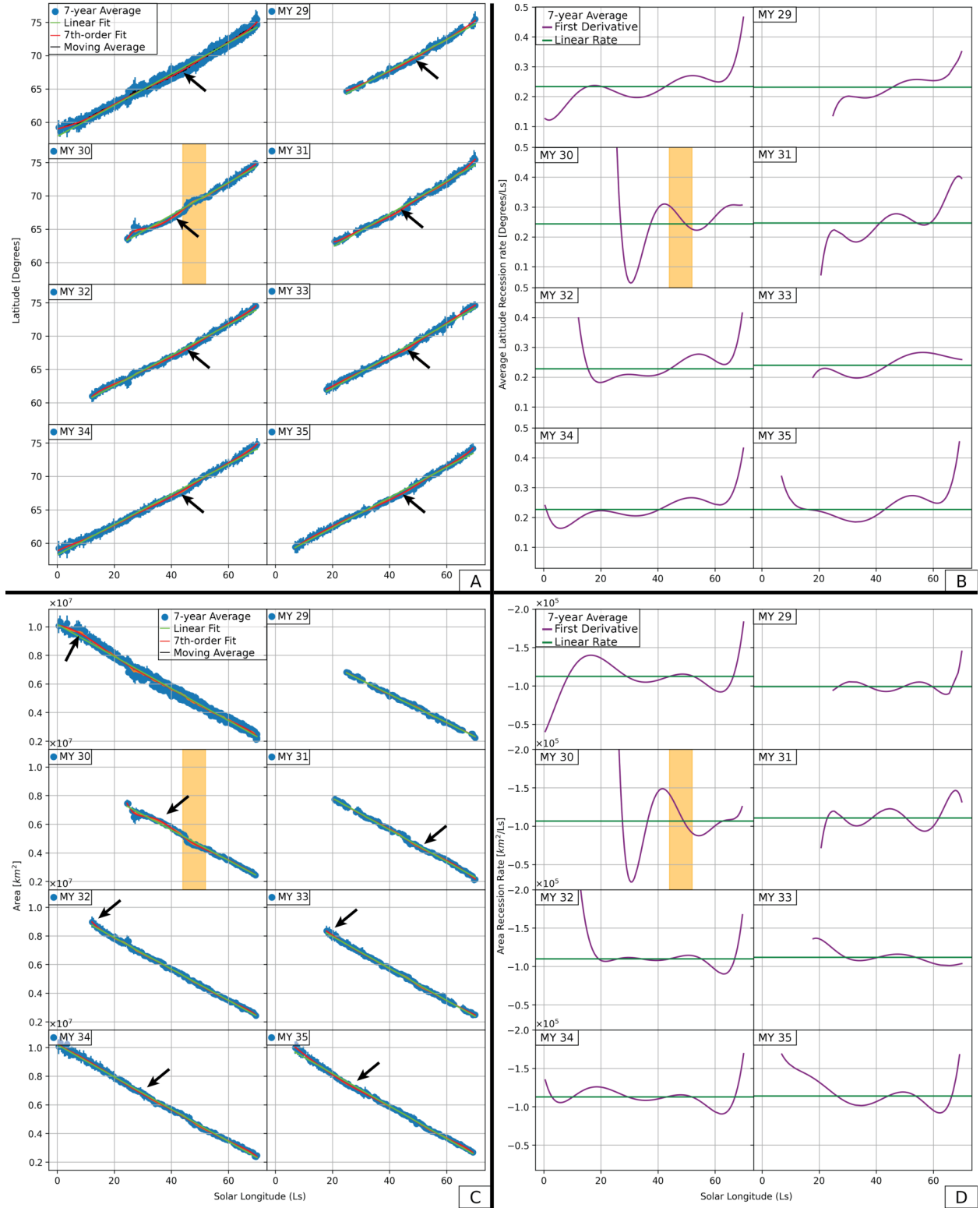


Figure 2.6: Fits for the seven-year average and each year individually with the yellow bars marking the sudden latitude increasing event during MY 30. A: Latitude recession of the NPSC axis with various fits. B: Latitude recession rate (green) and the first derivative of the 7th-order polynomial (purple). C: NPSC area extent with same fits. D: NPSC area recession rate (green) and the derivative of the 7th-order polynomial (purple). Black arrows point to the mid-season deviation from the linear fit.

2.3.2 Sudden Latitude Increase Events

We find a sudden latitude increase during MY 30 from $L_s = 44.4^\circ$ to $L_s = 45.7^\circ$. In this interval, the latitude rapidly increases by 0.8° (orange shading in Figure 2.6). Afterward, the cap size returns to the linear fit by $L_s = 51.5^\circ$, but the recession rate remains significantly slower and extends slightly further than the linear fit until $L_s = 61.7^\circ$. We also find that the recession rate is approximately 56% faster after the event, with the linear recession rate increasing from 0.18° per L_s before $L_s = 44.4^\circ$ to 0.28° per L_s after $L_s = 51.5^\circ$ (Figure 2.7a). We notice this coincides temporally with dust storms moving outward from the NPSC just before the increase (Figure 2.7b).

We believe there was an earlier increasing event; however, due to the low number of observations from $L_s = 24.5^\circ$ to $L_s = 26.3^\circ$, we are unable to observe the sudden increase. We find that from $L_s = 26.3^\circ$ to $L_s = 29.5^\circ$, the recession rate is significantly slower than average, qualitatively similar to the $L_s = 44^\circ$ event, and there is major dust activity from $L_s = 24.5^\circ$ to $L_s = 26.3$. The similarities between the periods suggest that both events are driven by similar processes. Additionally, this pattern does not recur in other MYs, nor do the storms appear at the same intensity.

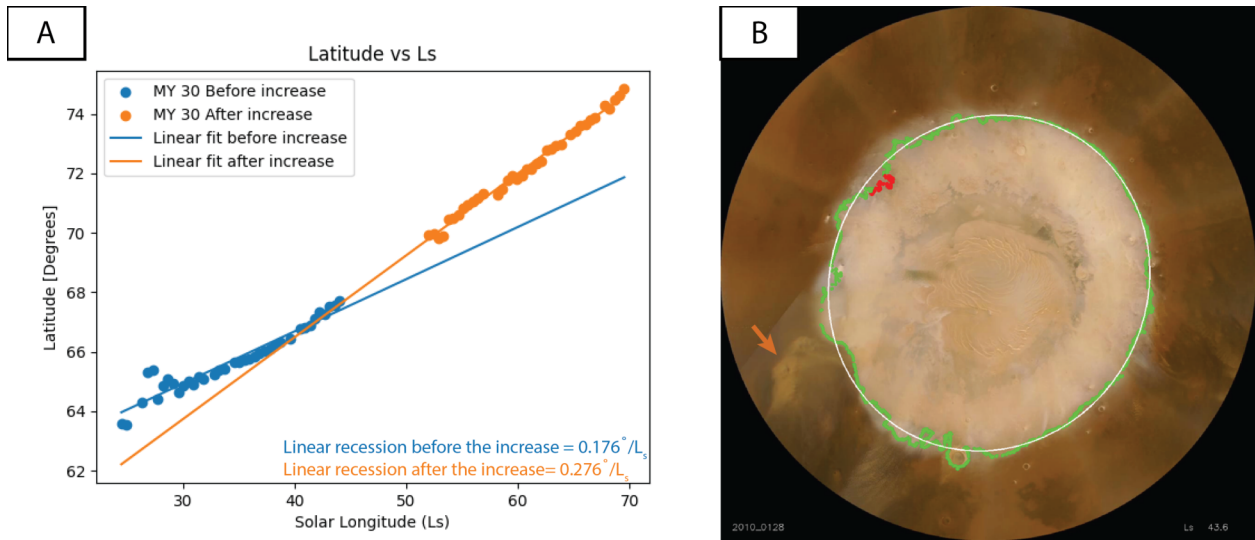


Figure 2.7: MY 30 observations A: Latitude vs L_s for MY 30 before the sudden increase at $L_s = 45^\circ$ (blue points and line) and after the increase (orange points and line). B: A dust storm (orange arrow) occurring during MY 30 at $L_s = 43.6^\circ$ that we hypothesize contributed to the enhanced sublimation.

2.3.3 Post-Global Dust Storm Years

For the two MYs that follow global storms, MY 29 and MY 35, the area and latitudinal extent of the seasonal caps deviate significantly from the MYs that do not follow dust storms. MY 29's seasonal cap tracks smaller and at a slower rate until $L_s = 45^\circ$ compared to other MYs (Figure 2.6c; Tables 2.2 & 2.3). MY 35's cap is slightly larger and recedes slightly faster throughout the observational period.

2.4 Discussion

2.4.1 Seasonal Cap Retreat Acceleration

Considering the general annual trends, we look at the broad annual steepening of the cap boundary curve (Figure 2.6b) and discuss the potential causes. Multiple factors affect the stable size of the NPSC and the seasonal recession rate. Repeatable conditions include the increasing solar elevation angle with L_s and the properties of the soil, including albedo and thermal inertia. Since the solar elevation angle should increase fastest early in the season (at $L_s = 0^\circ$, near the vernal equinox) and slow over the season (to the summer solstice at $L_s = 90^\circ$), one might expect the curve to be steepest early in the season and then decelerate. However, we observe the opposite, raising questions about what causes this acceleration. We present several possible explanations.

To explain this discrepancy, it would be reasonable to suggest that the surface albedo or thermal inertia of the soil changes by latitude, especially with higher ice concentrations farther northward; however, Putzig and Mellon (2007) only find evidence of strong thermal inertia changes at the highest latitudes of 75° and above, where the north polar residual cap resides. Similarly, surface albedo does not follow any progressive trend with latitude.

Another reasonable suggestion is the interplay between the seasonal CO_2 and water-ice. During early spring, the cap is dominated by CO_2 ice, and during mid-spring, it becomes partially covered by water-ice (Appéré et al. 2011). Interestingly, during late spring, large patches of CO_2 reappear

over the seasonal cap well within the cap's latitude (Appéré et al. 2011). It is suggested that water vapour is transported to higher latitudes via polar cells and/or baroclinic waves and condenses over the seasonal CO₂ as an optically thick but physically thin water-ice layer (Appéré et al. 2011). Then during late spring, the upper water-ice layer fractures and exposes the underlying cold CO₂ (Titus et al. 2020). We believe that this cannot directly affect the cap recession as the reappearance of CO₂ is well within the cap's edge and the edge of the cap is fully made up of water-ice. However, they may indirectly affect the cap recession rate due to increased surface wind activity.

Another possible explanation is a change in the sublimation temperature due to pressure changes with elevation or season. Piqueux et al. (2015b) found that the CO₂ sublimation temperature increases with pressure, and CO₂ ice is more stable at the bottom of the Hellas and Argyre basins in the southern polar region. However, the elevation decreases between 70° N to 75° N, which should increase the surface pressure and the stability of the CO₂ ice, although negligibly compared to the high-relief basins studied earlier. This won't explain the acceleration in retreat. Globally, the atmospheric pressure changes on seasonal scales; however, in the season of $L_{\odot}=50^{\circ}$ -70° the pressure curves from surface measurements are near local maxima, and between $L_{\odot}=0^{\circ}$ and $L_{\odot}=50^{\circ}$ the atmospheric pressure increases, which should increase stability and slow retreat. Thus, the acceleration in retreat cannot be explained by pressure changes.

Instead of soil, pressure, or H₂O blanketing effects, we believe that the increasing rate of recession can be explained more easily with atmospheric effects. In particular, sublimation enhancement caused by near-surface katabatic winds can explain the acceleration at higher latitudes. Spiga and Forget (2009) demonstrate that katabatic winds transfer momentum and energy to the surface, and Smith et al. (2013) show that up to 15% of the total energy budget to the surface can be caused by katabatic winds. Smith et al. (2015) and Smith and Spiga (2018) shows the seasonality of the highest velocity katabatic wind tracks in locations within the margin of the seasonal cap. Considering thermal contrasts between the exposed bedrock with low albedo and the frozen CO₂, Smith et al. (2015) suggest that the velocity of the katabatic winds near the seasonal cap boundary would be enhanced by a “sea breeze effect”. In addition, the interplay between water and CO₂ ice

may result in faster near-surface katabatic surface winds present near the edge of the cap. As the underlying CO₂ ice is exposed at higher latitudes, which is cooler than water-ice and sublimates faster, this will create a sublimation flow away from the pole. These enhancements may combine with winds originating near the cap's edge, resulting in strong winds near the cap's edge that transfer energy and momentum into the surface, accelerating sublimation and NPSC recession.

2.4.2 Sudden Latitude Increasing Event

In MY 30, we identify a rapid acceleration of retreat at $L_s = 45^\circ$ (Figure 2.7). Before the event, the recession rate is not substantially different from the annual average, but it jumps to over 0.3° per L_s briefly, marking the highest midseason retreat rate for all years. Because this is a one-off event, not repeated in other years, solar zenith angle, surface albedo, and thermal inertia cannot explain this acceleration. Thus, the atmosphere must be driving the acceleration.

In this scenario, before the latitude increases, the seasonal cap closely tracks the rising Sun (Figure 2.8). Then, we observe a significant retreat of the boundaries of the seasonal cap by 0.8° latitude. This coincides temporally with regional dust storms emanating from higher latitudes (Figure 2.7). We interpret this to indicate strong and short-lived near-surface winds coming off the NPLD that can strip surface materials and transfer energy to the surface (Figures 2.7b, 2.8b). These winds likely enhance sublimation in a similar way to the scenario later in the season. After the short acceleration, the recession rate slows to below average until the solar zenith angle catches up to the outer boundaries, where recession resumes at the late-season faster-than-average rate caused by enhancement from off-polar winds (Figures 2.8c-d) (Smith and Spiga, 2018).

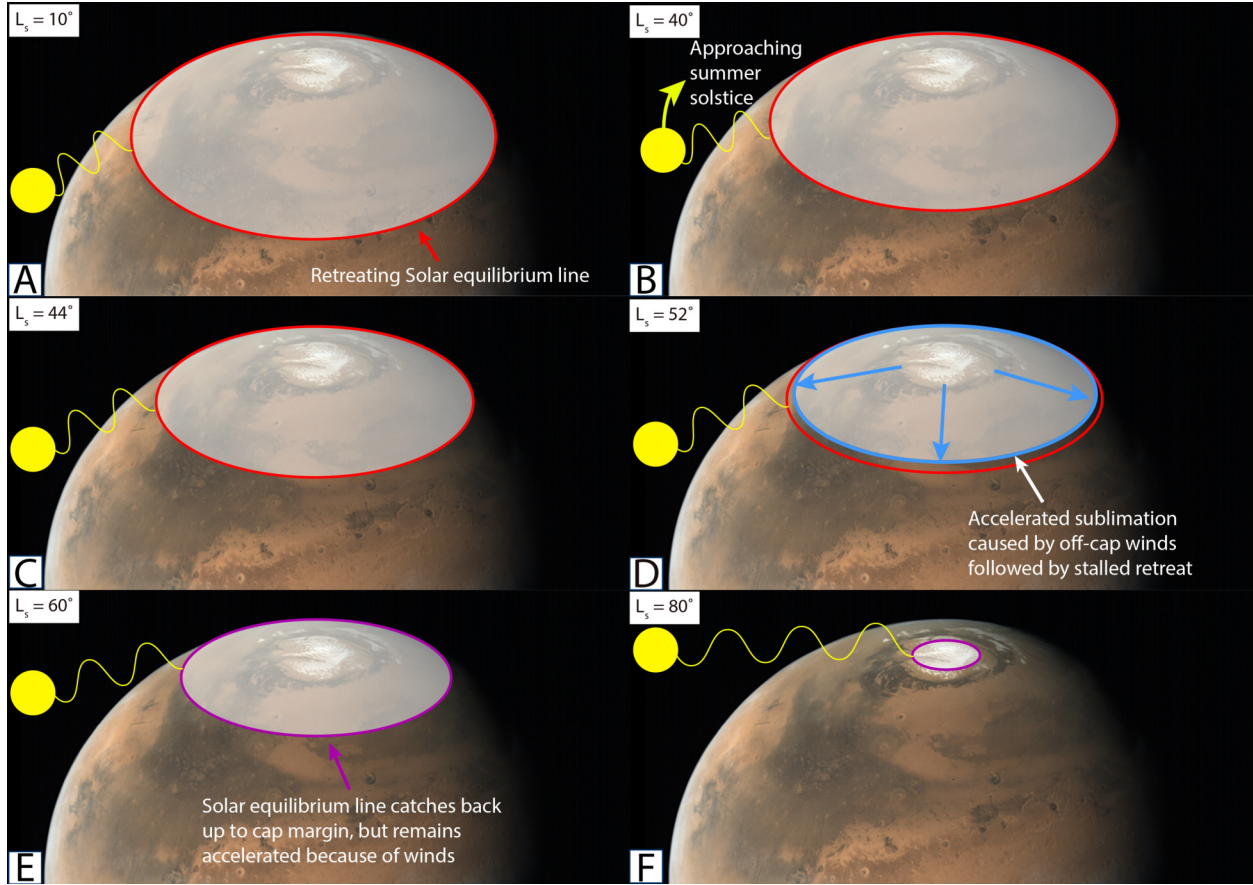


Figure 2.8: Heuristic model to explain sudden latitude increase in MY 30. The white-filled ellipse represents the seasonal polar cap, the yellow-filled circle represents the Sun, the yellow line radiating outwards from the Sun represents the solar radiation reaching the surface, and the red outlined/ white-filled ellipse represents the position of the sun after various L_s . From $L_s = 10^\circ$ to $L_s = 40^\circ$, the edge of the cap closely follows the maximum stable latitude of the Sun (Red Ellipse), which is mainly controlled by the solar elevation angle. Around $L_s = 44^\circ$, the NPSC shrinks (Blue Ellipse) well within the maximum stable latitude. Until $L_s = 52^\circ$, recession halts until the maximum stable latitude reaches the cap's edge, and around this time off-cap winds (Blue Arrows) are present on the surface. Once the maximum stable latitude reaches the cap's edge (Purple Ellipse), recession resumes; however, at a faster rate because the solar-driven recession is enhanced by off-cap surface winds.

2.4.3 Consequences of the Global Dust Storms

Two MYs exhibit very different behaviour than the seven-year average: MY 29 and MY 35 (Tables 2.2, 2.3). The NPSC in MY 29 is much smaller than average, with a much higher y-intercept in latitude, and recedes at a lower rate, whereas MY 35 is larger than average, albeit with a y-intercept close to the median of seven years. These seasonal caps follow global dust storms in

MY 28 and MY 34. We hypothesize that the abnormal behaviour of the seasonal caps in MY 29 and 35 is related to the GDS events.

The MY 28 GDS first appears at $L_s = 262^\circ$ (Wolkenberg et al. 2020), just as the NPSC reaches its maximum extent (Piqueux et al. 2015b), and mostly dies out by $L_s = 322^\circ$ (Wolkenberg et al. 2020). In contrast, the MY 34 global storm first appears much earlier, around $L_s = 185^\circ$ (Wolkenberg et al. 2020), before the northern seasonal cap begins to form (Piqueux et al. 2015b), and quickly dies out by $L_s = 220^\circ$ (Wolkenberg et al. 2020), about midway through the formation period (Piqueux et al. 2015b).

Piqueux et al. (2015b) postulated that the polar vortex is displaced during the MY 28 GDS. This would lead some regions of the cap to sublimate faster than others and faster than the interannual average, resulting in a smaller cap during MY 29; however, we are unable to observe the polar vortex using MARCI. Another possible explanation is as the dust storm slowly dies out, the airborne dust could fall onto the seasonal cap, lowering the surface albedo. The darker cap would sublimate faster due to higher insolation, resulting in a smaller MY 29 seasonal cap at the start of our observation period. The final explanation could be a combination of the previous two. Until mid-storm, the displaced polar vortex causes parts of the seasonal cap to sublimate faster than normal. Then as the storm dies out, dust is deposited onto the surface, lowering the surface albedo and the cap sublimates faster than average until the start of our observations.

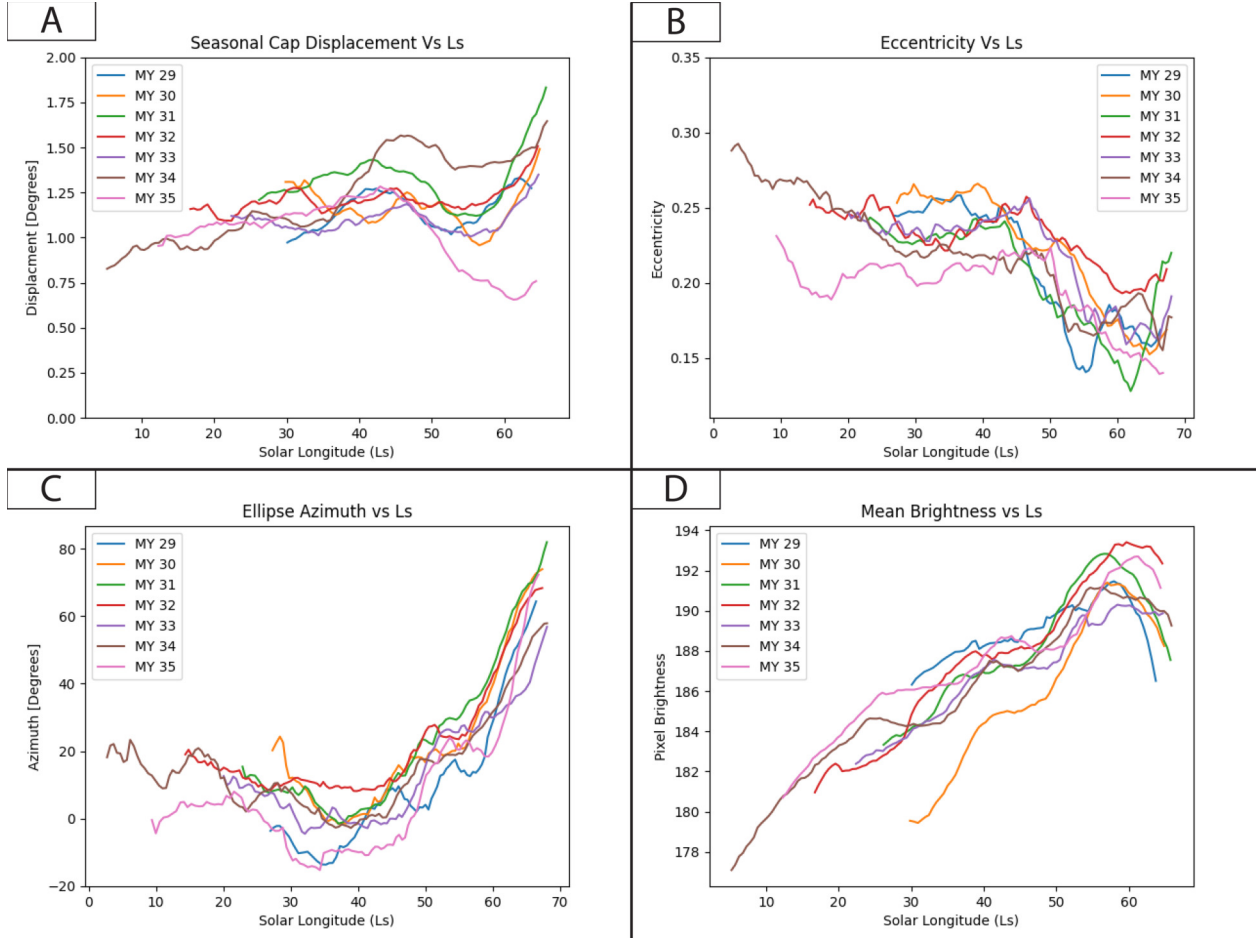


Figure 2.9: Parameters to test a moving polar vortex’s effect on NPSC retreat with a moving average of 10 points to enhance seasonal trends. A: The moving average displacement between the center of the filtered ellipse fit and the geographical North Pole vs L_s . B: The moving average eccentricity of the ellipse fit vs L_s . C: The moving average azimuth angle for the center of the ellipse fit. Negative values represent longitude values west of 0° E D: The moving average mean brightness of the outer 3° latitude of the seasonal cap vs L_s . MYs 29 and 35 do not strongly deviate from the interannual trends or even have smaller values than anticipated.

To evaluate the hypothesis put forth by Piqueux et al. (2015b), we measure three properties of the ellipse fit: the distance between the center of the ellipse and the geographic north pole (Figure 2.9a), the eccentricity of the ellipse (Figure 2.9b), and the azimuth of the ellipse (Figure 2.9c) for all MYs. With this, we test if the seasonal cap ellipse is shifted in MY 29 relative to the other years, possibly demonstrating the effects of a displaced polar vortex. If so, the azimuth, displacement, and eccentricity during MY 29 should deviate from the average. Contrary to prediction, we find that MY 29 has the smallest displacement at $L_s = 32^\circ$. Before $L_s = 40^\circ$, the azimuth angle varies

between 13° W and 6° W, indicating that the seasonal ice between 167° E and 174° E is closer to the pole. MY 29 has some of the highest eccentricity values before $L_s = 40^\circ$, then the eccentricity quickly decreases to below the multi-year norm. This could indicate that the cap is self-correcting, as regions less affected by the global storm are likely to extend further southward than those more affected. However, we do not identify any significant changes in the cap's extent and area during this period (Figures 2.5,2.6). These findings challenge the existing hypothesis suggesting that a smaller NPSC is attributable to a warmer atmosphere. However, due to the limitations of the MARCI—namely its inability to monitor the formation of the NPSC during the northern winter or directly measure surface temperatures—we are currently unable to test the idea of a warmer atmosphere.

To test the albedo-lowering method of increased sublimation, we use the V-channel of the MARCI mosaics to find the average brightness of the outermost 3° of the seasonal cap (Figure 2.9d) and find that MY 29 is slightly brighter, 2% brighter than the seasonal average until $L_s = 45^\circ$, again opposite of expectations. Interestingly, a similar trend is observed during the MY 21 GDS, which first appears around the same time of year as the MY 28 global storm (Cantor et al. 1998). After the storm, the seasonal cap recedes significantly more slowly and is smaller than average (Cantor et al. 1998), suggesting that late GDS events often lead to a smaller NPSC. That is, atmospheric warming may enhance sublimation before the onset of northern spring ($L_s = 0^\circ$) resulting in a smaller visible seasonal cap once the sun rises over the pole.

Now, considering MY 34, because the GDS quickly ends while the NPSC is still growing, there are more than 100° L_s dates between the end of the storm and the start of the recession. This gives time for the polar vortex to correct from the deviations and for the atmospheric dust levels to return to background levels.

Of note, previous work measures the average surface temperature during the MY 28 global storm and finds that it decreases by 10 K (Plesa et al. 2016) likely due to the major increase in the optical depth of the atmosphere and less solar radiation reaching the surface. Assuming similar conditions are present during the MY 34 storm when the seasonal cap is forming, we hypothesize

that the cooler surface at the time of NPSC formation would enhance the formation rate of the cap, leaving a thicker and broader cap during MY 35. Due to the lowered temperature, cold trapping is enhanced, which condenses more water and CO₂ than the normal rate, resulting in a larger overall cap. The larger cap would then have to sublimate faster early in the season from the lower latitudes, where the sun is more intense. This aligns with our observations of MY 35, which has the highest early-season recession rate.

2.5 Conclusion/Summary

We use daily MARCI mosaics and open-sourced computer vision library OpenCV to autonomously track the NPSC for MY 29 to MY 35 between $L_s = 10^\circ$ and $L_s = 70^\circ$. The backbone of our method uses HSV colour format, which decouples important properties of colour into their channels, allowing us to outline the cap while ignoring problematic water-ice clouds. Using the outline, we fitted an ellipse of best fit and recorded the average axis in latitude and the area of the ellipse. We characterized the recession curves using linear fit, a moving average, a 7th-order polynomial fit, and its first derivative. Our method was able to reproduce the work done by Calvin et al. (2015), who had manually tracked the seasonal cap for the same three years (MY 29-31). Small differences between the two methods are due to the difference in the shape of the best fit used. Due to the higher degrees of freedom, our ellipse fit was better suited for the non-circular shape of the mid-spring seasonal cap. Additionally, our method can be completed much more quickly, in about 2.5 minutes for all seven MYs. Because of the speed of the automation, we were able to process 99% of the observations, eliminating only those with unrecoverable artifacts. Future years will be analyzed very quickly.

Using the lines of best fit, we observe that most MYs follow similar trends to the seven-year average. In general terms, recession starts slowly and ends at a higher rate, agreeing with variable recession rates first identified by Kieffer and Titus (2001). We hypothesize that late-season, high-latitude near-surface winds coming from the NPLD deposit energy and momentum into the surface,

as first modelled by Spiga and Forget (2009). These would enhance sublimation.

We also find inter-annual variability in the recession rate. For example, we identify a rapid sublimation event in MY 30, followed by 6 Ls date when the recession slowed down. Similar to the annual increase in the recession, we believe that surface winds are responsible for the acceleration. In MY 29 and MY 35, the NPSC was smaller and larger than average, respectively. Previous work suggested that the seasonal caps in those years were affected by the MY 28 and MY 35 global dust storms. The contrasting behaviour from the two post-global dust storm years (between themselves and between other years) may be due to the difference in timing of the MY 28 and MY 34 storms. The late MY 28 storm may have displaced the polar vortex and/or lowered the surface albedo, resulting in a smaller cap during MY 29, and the early MY 34 storms may have enhanced the formation of the cap due to the lower surface temperatures, resulting in a larger MY 35 cap.

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Chapter 3

Tracking the South Polar Seasonal Cap Retreat of Mars Using Computer Vision

3.1 Motivation

One of the primary motivations behind this chapter is to build upon previous work that used daily observations from Mars Colour Imager (MARCI) to track the recession of the South Polar Seasonal Cap (SPSC) (Brown et al. 2010; Calvin et al. 2017). Calvin et al. (2017) manually outline the SPSC for MYs 28-31. However, their focus is not on determining the seasonal recession rates for those MYs; rather, they use the outline to demonstrate longitudinal variation in the recession and compare between MYs. Another study that uses MARCI observations is conducted by Brown et al. (2010), who combine MARCI and spectral observations from the Compact Reconnaissance Imaging System of Mars (CRISM) to map the seasonal CO₂ ice cap, comparing spectral and visual observations for MY 28. This chapter builds upon both works by monitoring the SPSC recession for multiple MYs, allowing for a more comprehensive analysis.

Furthermore, this chapter improves upon Chapter Two of this dissertation, which used a static colour detection algorithm. That static detector has a significant drawback when applied to the SPSC in its inability to account for dust storms that persist for several days at a time and cover

a substantial portion of the SPSC. Here, we improve upon the static detector by using a dynamic colour detector that can adapt to the dynamic southern polar region, allowing us to utilize more MARCI mosaics.

3.2 Methods

3.2.1 MARCI Dataset

This section uses the MARCI polar mosaics centred at the South Pole, spanning MYs 28-31, from $L_s = 178^\circ$ to $L_s = 345^\circ$. These mosaics are developed by Calvin et al. (2017) and are similar to the ones used in Chapter 2.

To determine the latitude conversion factor, we apply the identical method to those detailed in Chapter 2.2.1. We measure the distance between the geographic South Pole and various surface features in terms of both MARCI pixels and latitude degrees. This is facilitated using the JMARS software (See Section 2.2.1 for more detail). Through these measurements, we establish a consistent conversion factor of 0.072° latitude degrees per pixel or about 4.3 kilometers per pixel, which remains constant across this projection and is applied throughout the analysis in this chapter.

3.2.2 Initialization Loop

Similar to Chapter 2.2.2, the following sub-sections describe how we develop a dynamic algorithm to outline the SPSC and fit an ellipse of best fit using Python and OpenCV.

To initiate the tracking, we first convert the mosaics from Red, Green, and Blue (RGB) format to the Hue, Saturation, and Value (HSV) format. The HSV format separates crucial properties of colour such as saturation and brightness (Value) into their own channels, making it easier to distinguish between the SPSC, water-ice clouds, and unfrosted regions.

Table 3.1: Table presenting the initial upper and lower HSV bounds for the starting MARCI mosaic for each MY.

MY	Lower HSV Bound	Upper HSV Bound
28	(31,0,129)	(180,106,255)
29	(31,0,173)	(180,145,255)
30	(31,0,139)	(180,114,255)
31	(31,0,148)	(180,106,125)

Next, we empirically determine the upper and lower HSV bounds that best outline the SPSC for the starting mosaics for each MY (Table 3.1). Subsequently, we extract a binary image from the mosaics, assigning values of zero (black) for pixels outside the HSV bounds and 255 (white) for pixels within the bounds, same as Section 2.2.2. Then, we find all the contours in the binary image, representing regions where the pixel values transition from zero to 255, allowing the algorithm to identify the SPSC edge (Figure 3.1b). Many small contours/outlines are identified within the HSV bounds that do not match the extent of the SPSC and will negatively impact the fit of the ellipse fit (Figure 3.1b). To ignore these outlines, we combine the three large contours, by area, into a single contour that best matches the extent of the NPSC (Figure 3.1c). We chose to combine the three largest contours rather than the largest contour, which was used for Section 2.2.2, to reduce the impact of imaging artifacts such as mismatch in the colour channels and missing data (Colour fringes and black regions in Figures 3.1). We then apply an initial ellipse fit (Black Ellipse in Figure 3.1d). To refine the initial ellipse fit, we apply a simple filter to exclude any points that are greater than half of the semi-major axis distance from the center of the image (Red points in Figure 3.1d), resulting in a more accurate representation of the SPSC. Using the remaining points (Green points in Figure 3.1d), we refit the ellipse (White ellipse in Figure 3.1d) and record the semi-major, semi-minor, average axis, and the area, after converting the outputs from pixels to degrees and kilometers, as described in Section 2.2.3.

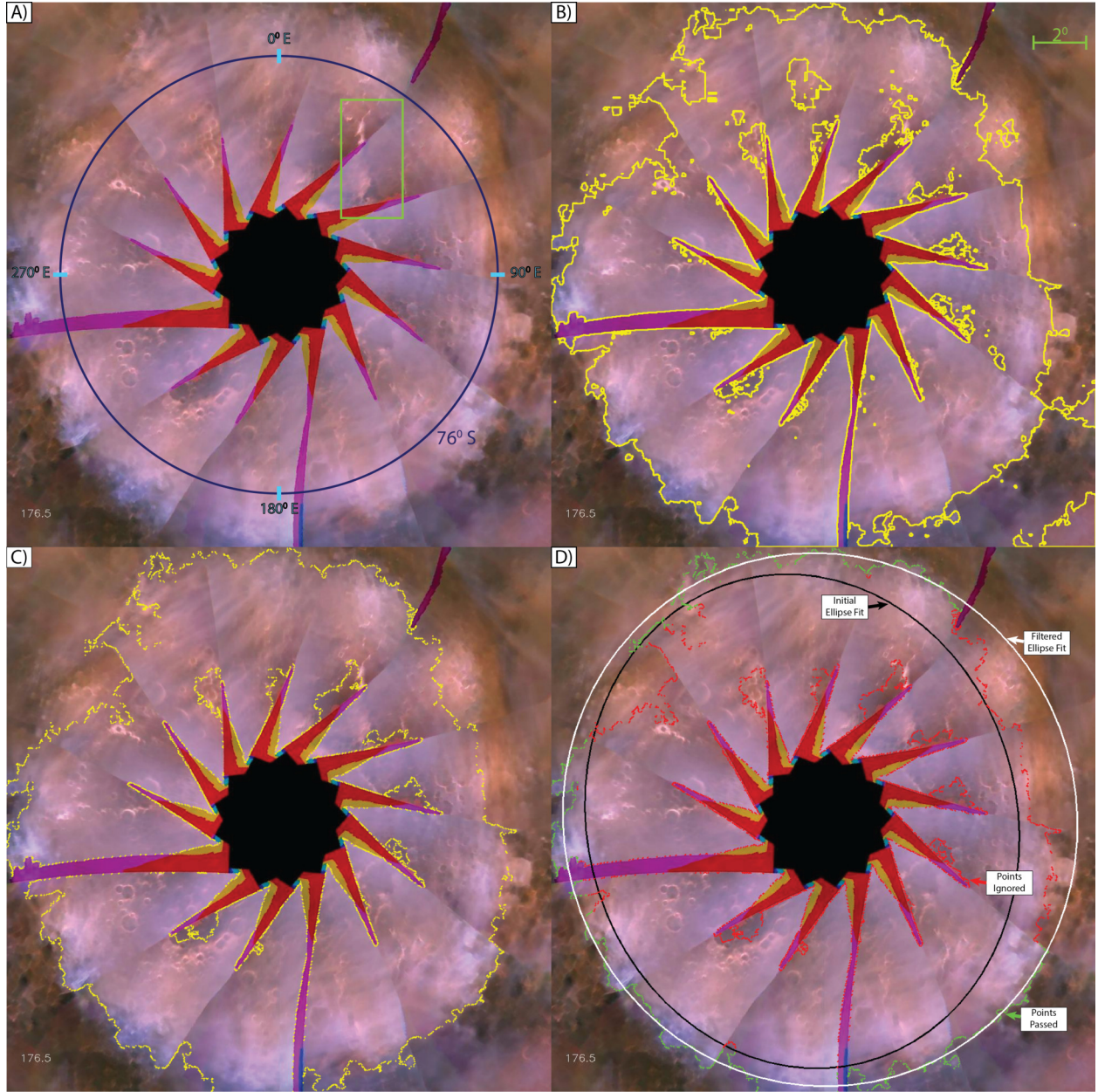


Figure 3.1: MY 28 $L_s = 176.5^\circ$: First MARCI mosaic for MY 28. A) Original MARCI mosaic displaying the south polar region. The green bounding box shows the extent of the Mountains of Mitchel seasonal cap (Figure 3.3) we consider for Section 2.3. B) MARCI mosaic with all contours (Yellow Line) overlaid delineating the boundaries where the automatic tracking detects the South Polar Seasonal Cap (SPSC). C) Combination of the three largest contours by area (Yellow Line), resulting in a contour that best aligns with the SPSC. D) Overlay of the initial ellipse fit (Black ellipse) and the final ellipse fit (White ellipse) on the mosaic. The red points are where the distance from the center is larger than half of the semi-major axes (excluded point) while the green points are points whose distance from the center of the ellipse is within half of the semi-major axis from the center (included points).

3.2.3 Dynamic Loop

In subsequent mosaics, we are using an elliptical mask with the same dimensions as the previous ellipse fit (The black region in Figure 3.2b). We divide the masked image into 2,500 boxes, each measuring six-by-six pixels (Red boxes in Figure 3.2b), to determine the average H, S, and V values of the new mosaic, carefully excluding any boxes containing black pixels. The lower S and V bounds are set at 60% of the mean value (lower would include other features such as water-ice clouds and higher would exclude some SPSC), while the upper bounds are set at 30% larger than the mean—a range determined through empirical analysis.

Next, we proceed with a larger ellipse mask, 25% larger than the mask in the previous time step (Figure 3.2c), and use a similar procedure as the initialization step: 1) converting the RGB frame to HSV, and 2) identifying all contours within the binary image. The presence of numerous small contours presents challenges for achieving an accurate initial fit (Figure 3.2d). To address this issue, we selectively consider contours with an area equal to or exceeding 25% of the maximum contour size. This serves as a basis for obtaining the initial ellipse fit; the 25% threshold is determined through empirical analysis. Similar to the initialization step, contour points fail to outline the SPSC. To remedy this, we employ a straightforward filtering algorithm that eliminates any contour points whose distance from the center of the initial ellipse fit is 25% greater than the semi-major axis (Figure 3.2f); this eliminates distant outliers and prevents them from affecting the ellipse of the bulk SPSC. Using the remaining points, we derive the filtered ellipse fit and record the same ellipse parameters as in the initialization step.

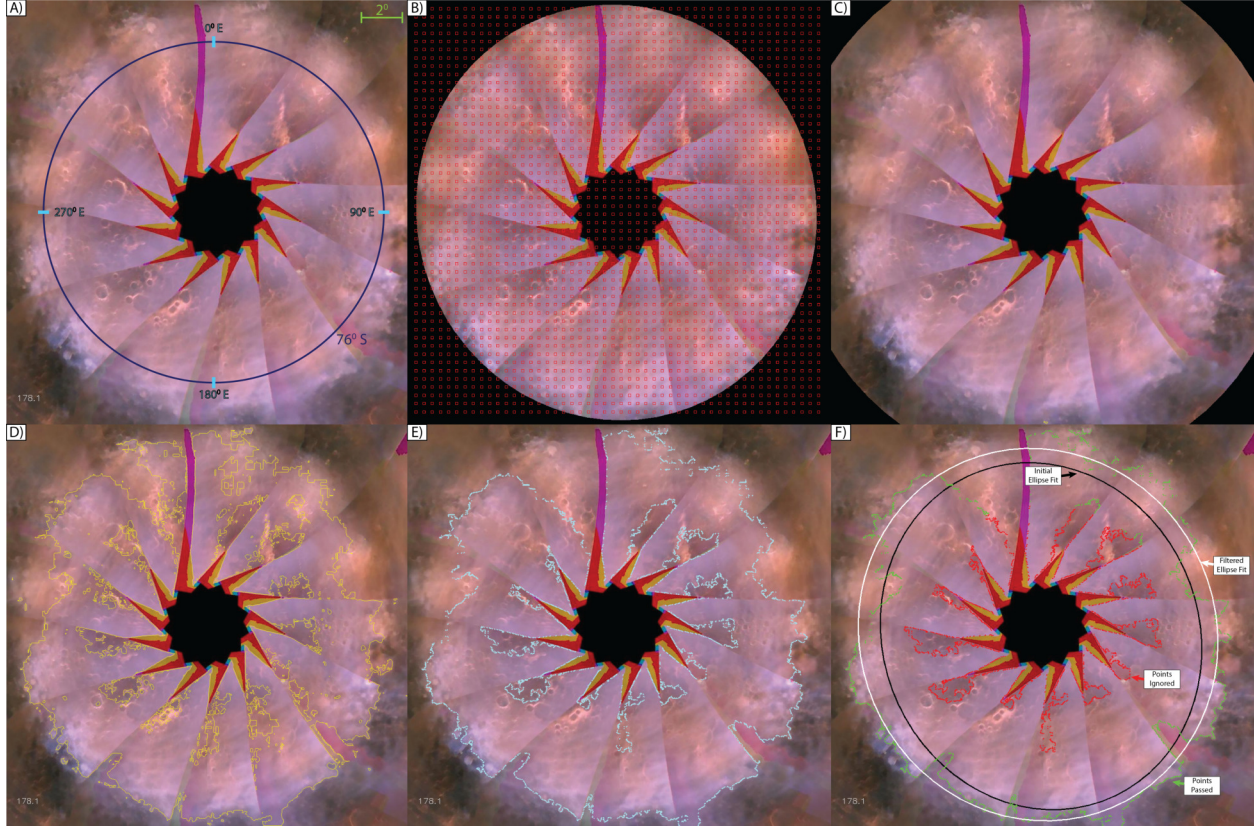


Figure 3.2: MY 28 $L_s = 178.1^\circ$ MARCI mosaic. A) Original MARCI mosaic B) MARCI mosaic with the ellipse mask derived from the previous step (initialization step) and the red boxes indicate the sampled regions for obtaining the average H, S, and V values. C) MARCI mosaic showing the 25% larger elliptical mask to ensure that all of the SPSC is permitted. D) MARCI mosaic with contours overlaid (Yellow Lines). Many small contours (Black Arrow) appear throughout the image making it difficult to outline the SPSC. E) MARCI mosaic highlighting the exclusion of contours that are less than 25% of the maximum area (Blue Lines). F) MARCI mosaic displaying the initial ellipse fit (Black Ellipse) and the filtered ellipse fit (White Ellipse), along with the ignored points (Red Points, mostly to exclude processing artifacts) and the points used to get to the filtered ellipse fit (Green Points)

3.2.4 Mountains of Mitchel Seasonal Cap Autonomous Tracking

As the SPSC undergoes sublimation, it breaks up into smaller fragments, producing two prominent 1s: the Mountains of Mitchel (MM) seasonal cap, primarily composed of CO_2 , and the late summer outlier, primarily composed of water-ice (Calvin et al. 2017). The MM seasonal cap, located approximately at 72° S , 30° E , was first observed by (Mitchel 1846). The MM seasonal cap exhibits significantly higher brightness than its surroundings and has been documented to display seasonal variability, albeit on a smaller scale than the SPSC (Bonev et al. 2002). The MM seasonal

cap is present at an elevated region, approximately 200 meters above the surrounding area. This region retains its seasonal frost for longer because the sublimation temperature decreases with elevation, higher elevations have a lower sublimation point. The frost outlier can account for a significant percentage of the SPSC, variability in the sublimation of the outlier can be important in explaining the variability of the SPSC.

We track the MM seasonal cap outlier from $L_s = 250^\circ$ to $L_s = 280^\circ$, observing its separation from the main SPSC until complete sublimation. To accomplish this, we use a static HSV colour detector employed in Chapter 2.2.2. Through empirical analysis, we determine the optimal lower and upper HSV bounds as (0,0,146) and (180,255,255), respectively. Before obtaining the binary image, we crop a rectangular box measuring 110 by 201 pixels centred over the MM seasonal cap at the pixel coordinates (655, 325; Figure 3.1a). The static colour detector performs well due to the limited L_s range and monitoring of a cropped region of the mosaic. The limited L_s range ensures consistent colour within the cropped region, while the cropped region effectively minimizes the influence of localized dust storms, which makes tracking the SPSC difficult. The static method removes the need for dynamic parameter tuning, which provides a more streamlined approach. Then, we identify the largest contour, in terms of area, that provides the most accurate outline of the MM seasonal cap (Figure 3.3). Due to the irregular shape of the outlier, a shape of best fit such as an ellipse is not sufficient to accurately characterize the outlier. For this reason, we record the area of the contour, which is then converted from pixel count to kilometers using the established conversion factor and the following equations.

$$S_{\text{Latitude}} = \Delta \text{Latitude} \cdot r \quad (3.1)$$

$$S_{\text{Longitude}} = \Delta \text{Longitude} \cdot r \quad (3.2)$$

Where $S_{\text{longitude}}$ and S_{latitude} is the arclength, Δs is the total latitude or longitude spanned by the rectangular crop, and R_r is the polar radius of Mars at 3373.3 kilometers (Haberle, 2015).

Using the equations above and the area of the rectangle, we can easily show the following to get the area of the MM seasonal cap in square kilometers:

$$A_{km} = \frac{r^2}{\alpha} \cdot \Delta\text{Pixels}_x \cdot \Delta\text{Pixels}_y \quad (3.3)$$

Where A_{km} is the total area, ΔPixels_x is the length of the crop along the short size (width of the image in Figure 3.3), ΔPixels_y is the length of the long side (height of the image in Figure 3.3), and α is the latitude conversion factor of 0.072° per pixel.

Using this relationship we can show the following:

$$\frac{\frac{Pixel}{MM} A}{\frac{Pixel}{Total} A} = \frac{\frac{km}{MM} A}{\frac{km}{Total} A} \rightarrow \frac{km}{MM} A = \frac{\frac{Pixel}{MM} A}{\frac{Pixel}{Total} A} \cdot \frac{km}{Total} A \quad (3.4)$$

Where $\frac{km}{Total} A$ is the total area in square kilometers, $\frac{Pixel}{Total} A$ is the total area in pixel squared, $\frac{Pixel}{MM} A$ is the area of the outline, the MM seasonal cap, that is found from OpenCV, and $\frac{km}{MM} A$ is the area of the outlier.

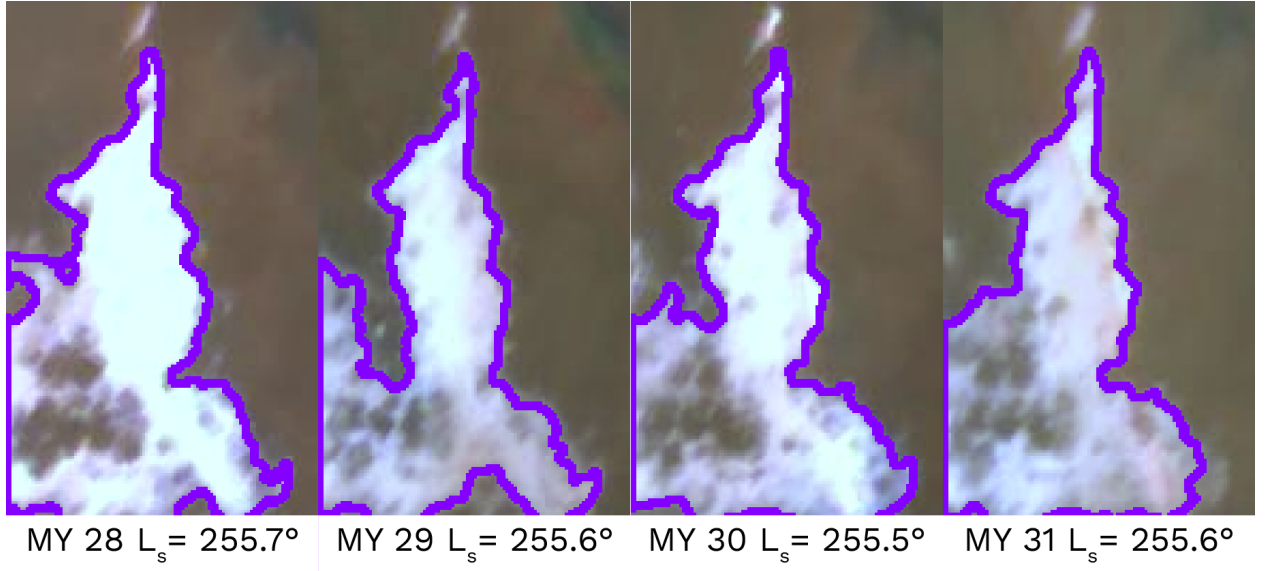


Figure 3.3: MARCI crop (bounding box can be seen in Figure 3.1a), approximately 2° by 4° , showing the Mountains of Mitchel seasonal cap and the automated outline (Purple Outline) for different MYs 28 – 31

3.2.5 Cleaning the MARCI Dataset

There are several instances where the ellipse fit fails to accurately match the SPSC edge (Figure 3.4a), and mosaics have large gores, which are regions with no data (Figure 3.4b). To tackle this issue of gores, we conduct an empirical examination of all mosaics before running the automated tracking code. We remove problematic mosaics from our dataset; specifically, MY 28 and MY 31 each contain one such mosaic, while MY 30 has three such mosaics, collectively representing approximately 1% of the total dataset.

To address the challenge of poor ellipse fits for the remaining mosaics, we empirically assess the final mosaics to ensure the accuracy of the fit. In a few cases, factors such as the MY 28 global dust storm or image artifacts result in a mismatch between the final ellipse fit and the SPSC. Usually, automated detection requires 1-2 mosaics to adjust its detection parameters (Figure 3.6c). Despite these challenges, the final dataset still incorporates over 95% of the total mosaics available.

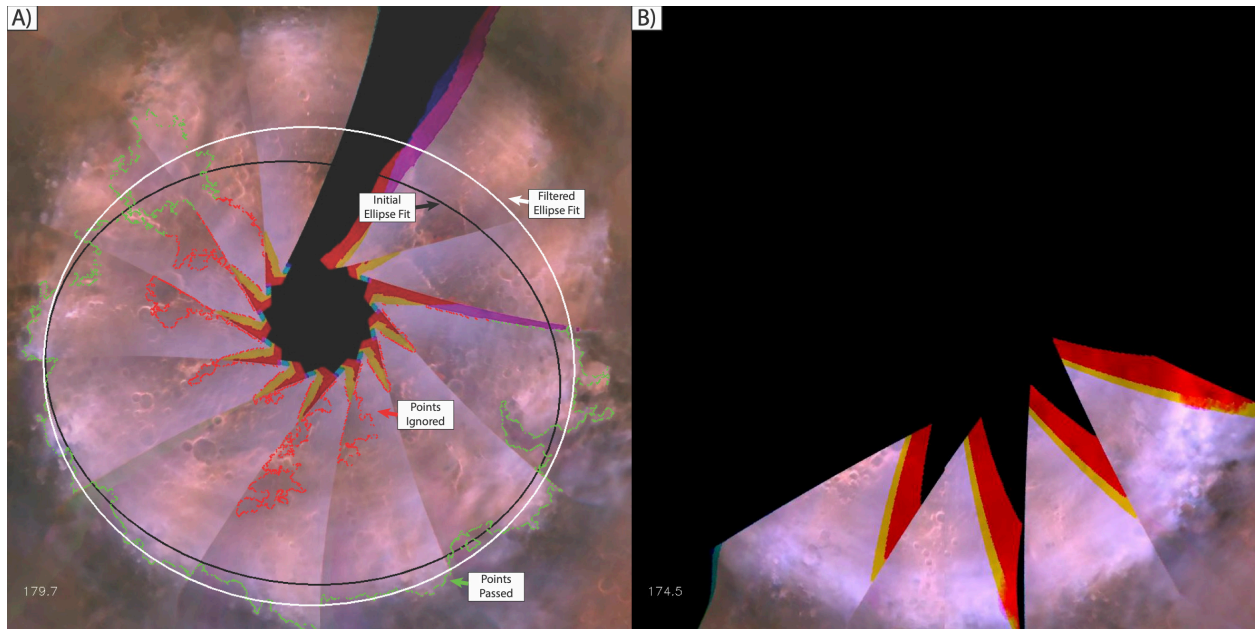


Figure 3.4: Problematic MARCI mosaics. A) MY 28 $L_s = 179.7^\circ$: An example of poor ellipse fit that is excluded from our final products. B) MY 30 $L_s = 174.5^\circ$: An example of a mosaic with large gores that make it impossible to determine the edge of the SPSC.

3.2.6 Validation

To validate our autonomous method, we compare the recession trends and the outlines from the automated method with previous studies that used spectral and manual methods. Specifically, we compare our findings with the outlines from Calvin et al. (2017) at $L_s = 205^\circ$, $L_s = 235^\circ$, and $L_s = 265^\circ$ (Figures 3.5a-b). My outlines closely align with theirs, displaying minor discrepancies. In certain instances, outlines appear slightly larger (dark blue arrows in Figure 3.5a), while in other cases, they appear marginally smaller (light blue arrows in Figure 3.5a)—the variations are less than 0.3° latitude. Furthermore, we compare trends with those of Brown et al. (2010) and Piqueux et al. (2015b), both of whom used spectral observations to track the seasonal CO_2 ice cap. We find that our SPSC is on average slightly larger than theirs, with the largest differences of about one million square kilometers occurring before $L_s = 180^\circ$. Using the same ellipse area equation as utilized in Chapter 2.2.3, we find that our SPSC area during MY 28 is 8% larger than the spectral SPSC area observed by Brown et al. (2010) (Figure 3.5c-d). The difference can be attributed to two factors: 1) MARCI's inability to differentiate between seasonal CO_2 and water-ice, and 2) the disparity in pixel scales. Due to a higher sublimation temperature, water-ice demonstrates greater stability at lower latitudes compared to CO_2 for a given L_s date, and both possess similar HSV values. As a result, when detecting seasonal ice, our method categorizes both types as a single type, thereby giving a larger extent. The dataset from Brown et al. (2010) has a spatial resolution of 187 meters per pixel, which is approximately ten times finer than our MARCI mosaics. In the coarser MARCI images, a pixel at a given latitude and longitude boundary identified as containing seasonal ice might not be fully covered by it but may be partially covered. This is not the case for images with finer resolution than MARCI, which results in a smaller extent of the SPSC.

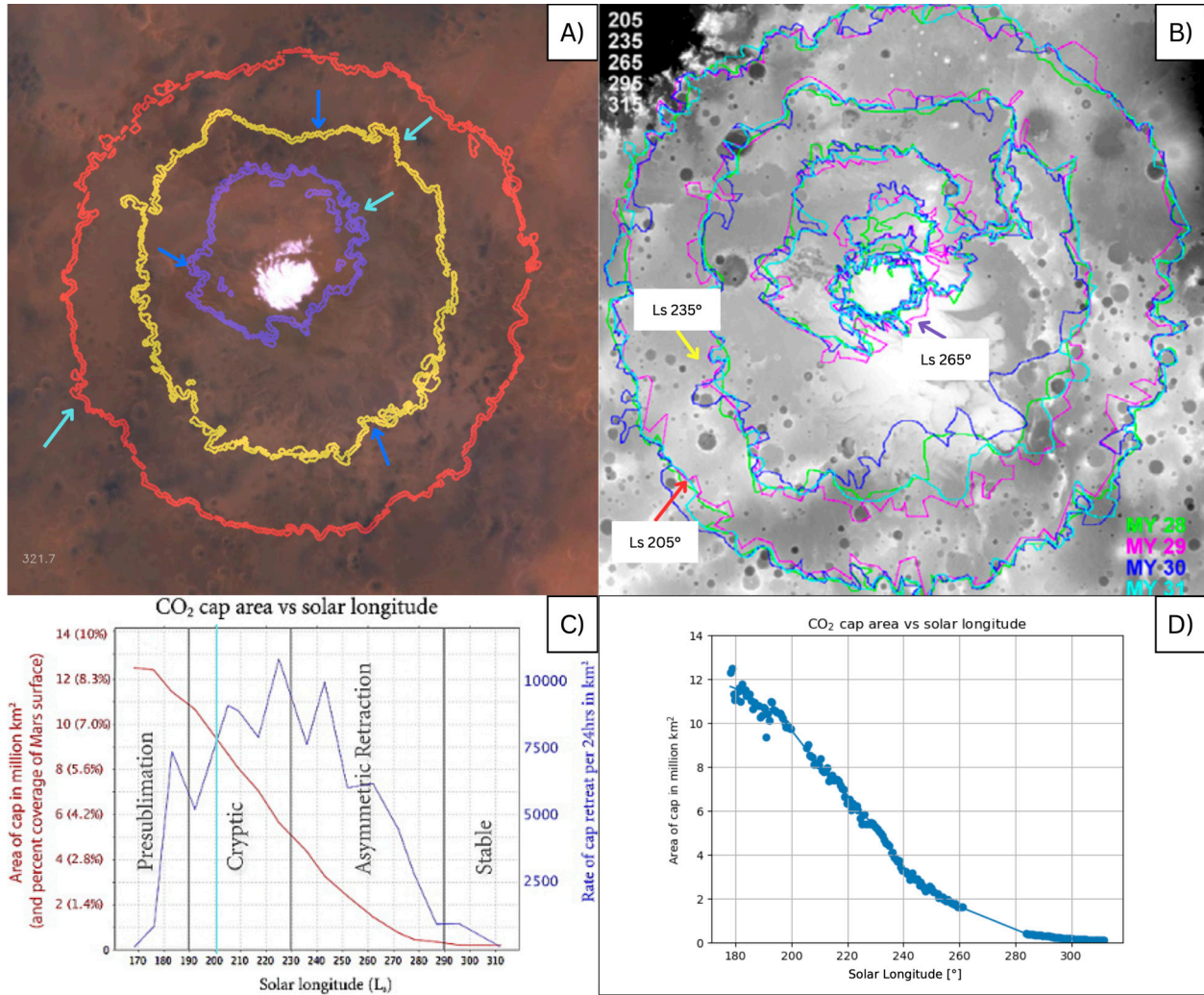


Figure 3.5: Comparing our results with Calvin et al. (2017) and Brown et al. (2010). A) Our outline for $L_s = 205.5^\circ$ (Red outline), $L_s = 235^\circ$ (Yellow outline), and $L_s = 265.2^\circ$ (Purple outline) overlaid on top of the MY 29 $L_s = 321.7^\circ$ MARCI mosaic. The dark blue arrows show regions where our outline is smaller than Calvin et al. (2017) and light blue arrows show regions where our outline is larger. B) Figure 3 from Calvin et al. (2017) with the arrows marking the L_s dates of the corresponding outlines found in 5A. C) Figure 6 from Brown et al. (2010) shows the CO_2 SPSC area vs L_s using spectral observations for MY 28 (Red line). D) Our area of the ellipse (Blue Points) and the non-parametric smoothing line (Solid Line) vs L_s date for MY 28. Both methods produce similar results; however, our SPSC area is about one million square kilometers larger than Brown et al. (2010).

3.2.7 Autonomous Performance

The autonomous detection analyzes over 900 MARCI mosaics within a span of 90 seconds. This is slower than the automated method discussed in Chapter 2.2; however, it offers improved adaptability to atmospheric processes such as dust storms occurring over the SPSC.

The autonomous method significantly outperforms manual methods in outlining the SPSC and extracting the SPSC recession in terms of processing speed and temporal resolution. The speed advantage comes from using a computerized method, which can process more images per second than manual methods. The temporal resolution advantage comes from using 90% of the MARCI mosaics (typically, mosaics are skipped to reduce computation time). The higher temporal data may provide new insights into the seasonal variability of SPSC as well as the MM seasonal cap. This detailed data can show us new patterns, like how small regional storms might speed up the melting of the SPSC. Also, by looking at data from different MYs, we might find unique yearly patterns, such as the sudden increase in latitude observed in MY 30, as discussed in Section 2.4.2.

During the MY 28 global dust storm, which spans from $L_s = 262^\circ$ to $L_s = 322^\circ$ (Wolkenberg et al. 2020), our autonomous method faces challenges at the start of the GDS. This is because the entire south polar region undergoes intense dust activity, obscuring the SPSC and resulting in a smaller outline (Figure 3.6b). However, the tracking swiftly adapts to the HSV bounds as the dust opacity diminishes and the SPSC begins to be visible again, around $L_s = 278.6^\circ$. Another important note is the discontinuity that occurs around $L_s = 260^\circ$, coinciding with the detachment of the MM seasonal cap from the main SPSC. This occurs because the automated method determines the MM feature to be too small to be considered part of the main SPSC.

3.2.8 Data Analysis

We divide our analysis into two distinct segments: before and after the $L_s = 260^\circ$ discontinuity (Figure 3.6). The precise date of the discontinuity by year can be found in Table 3.2. To extract sub-seasonal trends, we employ a nonparametric smoothing function (dashed lines in Figure 3.6) and its first derivative, allowing me to monitor periods of acceleration and deceleration in the latitude recession rates. This approach enables me to identify periods of accelerated recession and effectively minimize the overall noise in the data. In this method, the smoothing percentage is critical. This refers to the proportion of data we choose to uniformly ignore during non-parametric smoothing. We empirically determine the best parameters for each MY and segment (Table 3.2).

Segmentation is needed due to the large discontinuity arising from the separation of the MM seasonal cap. For the first derivative, we apply a smoothing percentage of 10%, a value determined through empirical analysis. Notably, because the latitude recession rate values are negative—due to negative latitudes—the more negative the values, the faster the SPSC is receding.

Table 3.2: Table showing the Discontinuity Dates and Smoothing Percentages used for non-parametric smoothing. Smoothing percentage presents the percentage of data ignored.

MY	Discontinuity Date	Smoothing Percentage (Before, After)
28	261.1	(5, 8)
29	263.4	(5, 8)
30	259.8	(6, 6)
31	259.7	(6, 8)

3.3 Results

3.3.1 SPSC Variability before the $L_s = 260^\circ$ Discontinuity

The latitudinal recession of the SPSC demonstrates a relatively linear trend with oscillations around the linear fit, revealing five notable periods of interest (Figure 3.6c). Table 3.3 provides a general behaviour of the SPSC during the five periods. In the initial period spanning from $L_s = 175.3^\circ$ to $L_s = 194.6^\circ$, the SPSC latitude is smaller than the linear fit and the latitudinal recession rate is relatively stable at approximately -0.2° latitude per L_s (Figure 3.6e), sublimating slower than the linear fit. Between $L_s = 196.4^\circ$ and $L_s = 215.4^\circ$, the SPSC latitude extends farther than the linear fit and the latitudinal recession rate generally accelerates; however, it is still sublimating slower than the linear fit. The next period, between $L_s = 215.4^\circ$ and $L_s = 234.5^\circ$, sees the SPSC latitude remaining larger than the linear fit but the difference between the linear fit and the SPSC is decreasing. Additionally, the latitudinal recession rate accelerates and remains faster than the linear fit. Most MYs (Figure 3.6e), except MY 31, reach their respective fastest latitudinal recession rate at $L_s = 234^\circ$. Between $L_s = 234.5^\circ$ and $L_s = 254.0^\circ$, the SPSC latitude is once again smaller than the linear fit, however, the difference is much smaller than in the previous period. During this time,

there is a general deceleration of the latitudinal recession rate; however, there is a brief accelerated recession, and the latitudinal recession rate reaches the second fastest rate at around $L_s = 252^\circ$. For the final period, from $L_s = 254.0^\circ$ to the discontinuity date, $L_s = 260$, the latitude of the SPSC edge continues to remain smaller than the linear fit (Figure 3.6c) and the latitudinal recession rate rapidly decelerates following the fastest recession at $L_s = 252^\circ$ (Figure 3.6e).

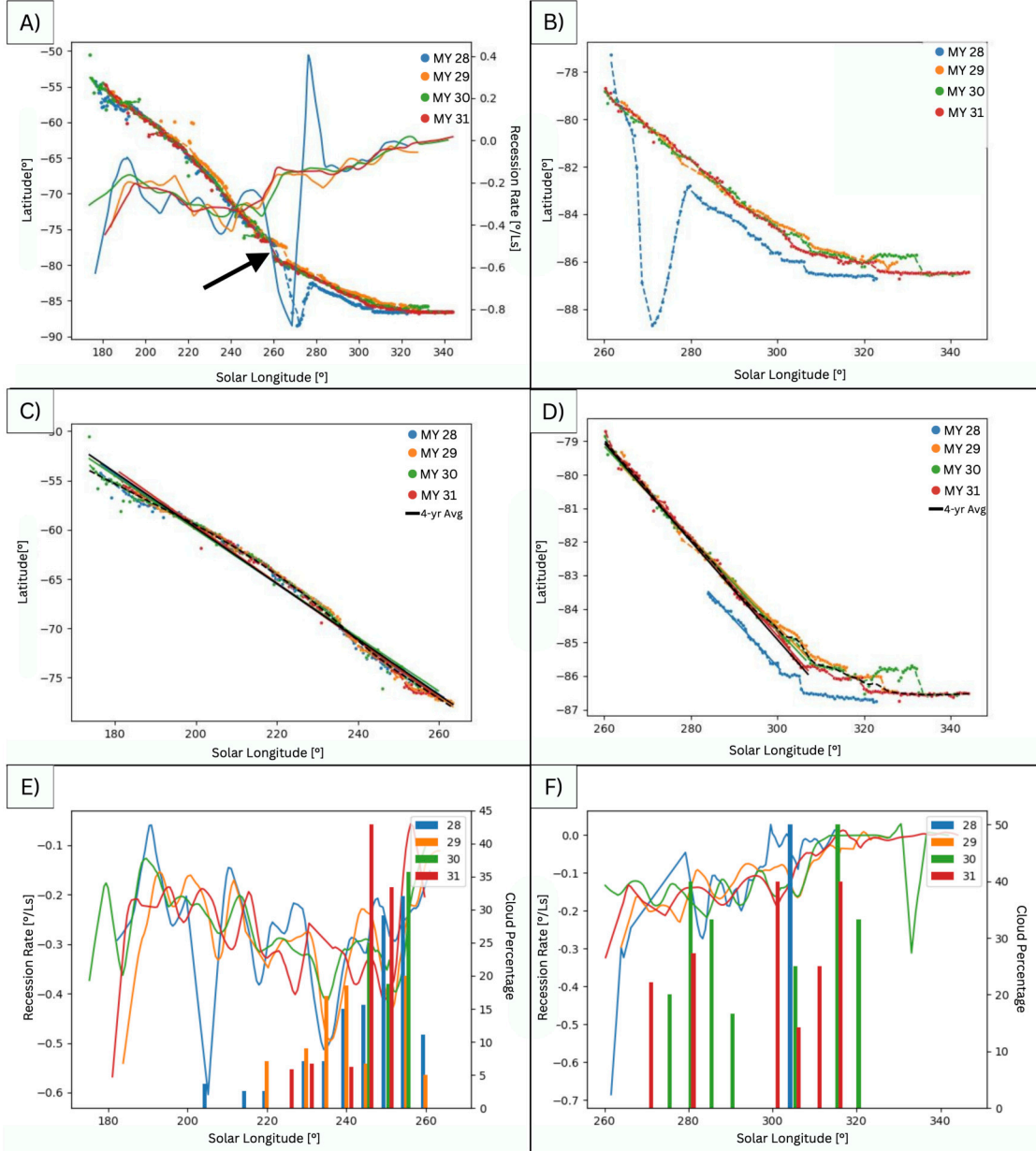


Figure 3.6: Output from the dynamic method showing the SPSC latitude vs L_s . A) SPSC extent and the seasonal recession rate (Solid Line) vs L_s for all four MYs with nonparametric smoothing (Dashed line). A discontinuity appears (marked by a black arrow) around $L_s = 260^\circ$ for all four years when the Mountains of Mitchel disconnect from the SPSC. In MY 28 there is an additional discontinuity due to the MY 28 global dust storm that obscures the surface. B) Similar to A but from $L_s = 260^\circ$ to $L_s = 320^\circ$, SPSC extent after the discontinuity, better shows the MY 28 global dust storm – the automated method quickly recovers after the onset of the storm, within 3° of L_s dates. C) SPSC extent with outliers removed; nonparametric smoothing (Dashed Line) and a line of best fit (Solid Line) before the $L_s = 260^\circ$ discontinuity. D) Similar to 6C SPSC extent after $L_s = 260^\circ$ discontinuity. E) Recession rate (Solid Line) before the $L_s = 260^\circ$ discontinuity and the cloud percentage (Bars) MYs 28-31 from Smith et al. (2015), binned every 5° of L_s date. During this period, the recession rate is highly variable. F) Similar to E but the recession rate (Solid Lines) and cloud percentages (Bars) after the $L_s = 260^\circ$ discontinuity.

Table 3.3: General observations of the five periods for the 4-year average and individual years. Latitudinal Recession Rate (RR) and Latitude (Lat) pertain to alternating rows.

MY	Period 1 ($L_s = 175.3^\circ$ - 194.6°)	Period 2 ($L_s = 194.6^\circ$ - 215.4°)	Period 3 ($L_s = 215.4^\circ$ - 234.5°)	Period 4 ($L_s = 234.4^\circ$ - 254.0°)	Period 5 ($L_s = 254.0^\circ$ - Discontinuity)
4Y AVG Lat	SPSC smaller than linear fit	SPSC larger than linear fit	SPSC larger, difference decreasing	SPSC smaller than linear fit	SPSC smaller than linear fit
4Y AVG RR	Stable at $-0.2^\circ/L_s$	Accelerates, slower than linear	Accelerating, fastest at $L_s = 235^\circ$	Decelerating, fastest at $L_s = 252^\circ$	Rapid deceleration to $0^\circ/L_s$
28 Lat	Smaller than other MYs	Larger than linear fit	Larger, difference decreasing	Smaller than linear fit	Smaller than linear fit
28 RR	Variable, fastest rate	Acceleration, slower than linear	Fastest rate at $L_s = 234^\circ$	Deceleration, slower between $L_s = 247^\circ$ - 250°	Rapid deceleration to $0^\circ/L_s$
29 Lat	Smaller than linear fit	Larger than linear fit	Larger, difference decreasing	Smaller than linear fit	Plateaus after $L_s = 254^\circ$ at $76^\circ S$
29 RR	Oscillates at $-0.22^\circ/L_s$	Acceleration, slower than linear	Fastest rate at $L_s = 235^\circ$	Deceleration, slower at $L_s = 241^\circ$ and $L_s = 245^\circ$ (local fastest recession rates at $L_s = 252^\circ$)	Rapid deceleration, faster than 4Y AVG
30 Lat	Smaller than linear fit	Larger than linear fit	Larger, difference decreasing	Smaller than linear fit	Stable at $76^\circ S$
30 RR	Follows 4Y AVG, slower than linear	Acceleration, slower than linear	Fastest rate at $L_s = 235^\circ$	Deceleration, varies more than 4Y AVG	Rapid deceleration after 254°
31 Lat	Smaller, larger than other years	Larger than linear fit	Larger, difference decreasing	Smaller than linear fit	Stable at $76^\circ S$ after $L_s = 253^\circ$
31 RR	Stable at $-0.23^\circ/L_s$	Acceleration, slower than linear	Fastest rate at $L_s = 230^\circ$, tied with MY30	Acceleration, second fastest at $L_s = 251^\circ$	Rapid deceleration, fastest rate

3.3.2 SPSC Variability after the $L_s = 260^\circ$ Discontinuity

During most MYs, except for MY 28, the initial latitude of the SPSC, around $L_s = 260^\circ$, is consistently nearly 79.87° S (Figure 3.6d). Before $L_s = 307^\circ$, the latitudinal recession of the SPSC proceeds linearly across all MYs, converging gradually towards the latitude of 86.1° S, the boundary of the residual cap. Notably, during this phase, the recession rate of the latitude exhibits minor variability and tends to exceed the rates predicted by the linear fits (Figure 3.6f). This variability is considerably less pronounced than the fluctuations observed in the period before the $L_s = 260^\circ$ discontinuity. Another discontinuity appears during late summer, around $L_s = 320^\circ$, which aligns with the sublimation of the late-summer outlier. This phenomenon arises when our automated method identifies this outlier as too small to be part of the principal cap. While the timing of this late-summer discontinuity varies significantly across MYs, the magnitude remains consistent at about 0.5° latitude. The latest occurrence of this discontinuity is during MY 30 at $L_s = 322^\circ$, and the earliest during MY 29 at $L_s = 319^\circ$.

During MY 28, the SPSC behaves more differently compared to other years. During the period after the MY 28 GDS event, the SPSC is discernibly smaller than average, approximately 3° latitude (Figure 3.6). Additionally, during MY 28, the SPSC reaches the latitude of the residual cap by $L_s = 307^\circ$, about 20° of L_s dates earlier than average. It is also noteworthy that the late-summer discontinuity appears at $L_s = 304^\circ$, again preceding the average by 20° of L_s dates. A significant characteristic of the SPSC during MY 28 is the accelerated sublimation rate, as evidenced by the steepest slope in the linear fits, alongside a notable increase in variability before $L_s = 307^\circ$.

3.3.3 Mountains of Mitchel Seasonal Cap Variability

In general, the MM seasonal cap first appears around $L_s = 252^\circ$ and then undergoes a mostly linear retreat, though with some variability, until $L_s = 275^\circ$, at which point it completely disappears (Figure 3.7a). Excluding observations during MY 28, the area of this feature between $L_s = 252^\circ$ and $L_s = 254.4^\circ$ initially exceeds the predictions of the linear fit. From $L_s = 258.9^\circ$ to $L_s = 275^\circ$,

the area of the MM outlier falls below the linear fit. Before $L_s = 264^\circ$, the rate at which the area recedes fluctuates around the linear fit, peaking at $L_s = 260.1^\circ$ and reaching the annual minimum at $L_s = 264.4^\circ$ (Figure 3.7b). This is followed by a phase of acceleration until $L_s = 267.6^\circ$, after which the recession rate steadily decreases towards zero by $L_s = 275^\circ$. Notably, during MY 31, the area recession rate shows less variability and remains relatively stable between $L_s = 259.2^\circ$ and $L_s = 266.2^\circ$, compared to other Martian years.

During MY 28, the behaviour of the MM seasonal cap is notably distinct from other years. Before $L_s = 260.3^\circ$, the starting date of the MY 28 GDS, the MM seasonal cap is significantly larger than in other years, though the area recession rate is similar to the average. After $L_s = 260.3^\circ$, the MM seasonal cap accelerates, leading to the complete sublimation of this outlier by $L_s = 268.1^\circ$ — approximately 12° of L_s dates earlier than the average. Additionally, the area recession rate exhibits less variability than in other years, with a relatively stable period between $L_s = 255.9^\circ$ and $L_s = 260.1^\circ$. A sharp acceleration is observed between $L_s = 260.1^\circ$ and $L_s = 262^\circ$, coinciding with the onset of the GDS.

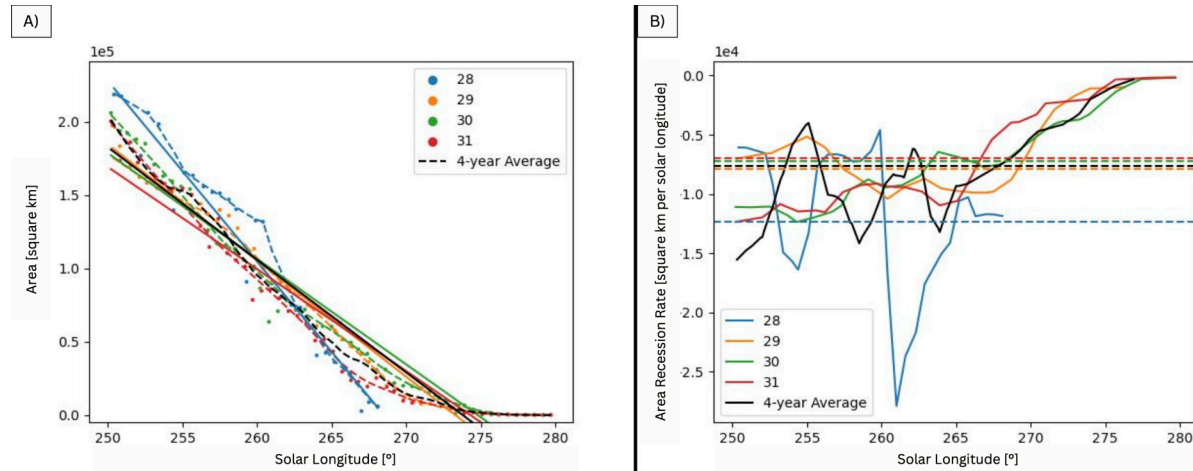


Figure 3.7: Mountains of Mitchel Area (in square kilometers) vs L_s with the nonparametric smoothing (Dashed Lines) and the line of best fit (Solid Lines) – area scaled by 10^5 . B) Area recession rates (in square kilometers per L_s (Solid Lines) vs the linear slopes (Dashed Lines) – scaled by 10^4 .

3.4 Discussion

3.4.1 General SPSC Variability

We find a positive correlation between the acceleration in the latitudinal recession rate and surface activity. This relationship is evident when aligning the cloud coverage data from Smith et al. (2015) with our recession rates (Figure 3.6e-f). Notably, before the $L_s = 260^\circ$ discontinuity, Smith et al. (2015) observe the annual maximum cloud sampling of 25% at all images at $L_s = 252^\circ$, coinciding with the second-fastest latitudinal recession rate, rather than the annual maximum at $L_s = 235^\circ$. During $L_s = 252^\circ$, the edge of the SPSC is proximate to the region characterized by steeper slopes, which allows for faster katabatic winds and thus more cloud formation (Smith et al. 2015). Smith et al. (2015) suggested that the presence of the SPSC on the steep slopes would enhance the existing katabatic winds with the sea breeze effect caused by circulation from colder to higher regions near the surface. It is also notable that cloud detections nearly double from 5% to 9% between $L_s = 230^\circ$ and $L_s = 239^\circ$, a percentage increase exceeding that between $L_s = 249^\circ$ and $L_s = 255^\circ$, the time of the second fastest recession rate.

We interpret this to mean that the latitudinal recession rate is influenced by fluctuations in surface activities, of which wind speed is the most important. During MY 31, the SPSC retreat lags by approximately $5^\circ L_s$ dates from the other years, with the annual fastest recession rate occurring around $L_s = 230^\circ$ —rather than $L_s = 235^\circ$. This lag could be attributed to the reduced surface activity, namely fewer trough clouds implying lower wind speed, as there is no major surface activity until $L_s = 245^\circ$.

3.4.2 Consequences of the MY 28 Global Dust Storm

Following the onset of the GDS around MY 28 $L_s = 260^\circ$, the unique behaviour of the SPSC can be attributed to the effects of the GDS (Figure 3.6). A similar trend is observed by Piqueux et al. (2015b) using spectral observations, where the SPSC is smaller during and after the GDS event.

Chapter 2.4.3 and Calvin et al. (2015) also observe analogous trends in the NPSC during MY 29, immediately following the GDS event. Both investigations measure the NPSC as being smaller than the interannual average and exhibiting slower recession rates. As mentioned in Chapter 2.4.3, the elevated atmospheric temperatures resulting from the GDS might have weakened the formation of the NPSC and accelerated the sublimation rates before their observational period.

Expanding on these findings, we postulate that the MY 28 GDS event might exert a comparable influence on the SPSC, resulting in an elevated atmospheric temperature during the event and subsequently a faster recession rate. A warmer atmosphere is observed during the event (Plesa et al. 2016), consistent with our observations of a smaller residual cap following the end of the GDS.

3.4.3 Seasonal Variability in the Mountains of Mitchel Seasonal Cap

The MM seasonal cap exhibits a significantly faster area recession rate during MY 28 compared to other years (Figure 3.7). This observation aligns with the findings from Calvin et al. (2017), who note a qualitatively accelerated sublimation of the outlier during MY 28; however, they do not present the overall area of the MM seasonal cap. A similar trend is observed during MY 25, another GDS year, during which the Mars Orbiter Camera observes that the MM outlier sublimates faster than in non-dust storm years (Bonev et al. 2002). We hypothesize that the faster sublimation of the outlier is due to warmer temperatures present during and following the MY 28 GDS.

One aspect of the MM outlier that remains puzzling is the disparity in the area of the outlier during MY 28 before the onset of the MY 28 GDS at $L_s < 260^\circ$. Before the MY 28 GDS, the MM seasonal cap outlier is approximately 20% larger in area and sublimates about 63% faster than the 4-year average (Figures 3.3, 3.7). This peculiar behaviour of the MM could serve as an indicator of an impending global dust storm. In observations of the larger SPSC, we find that larger caps sublimate faster than the seasonal average, which could enhance near-surface storms and merge small storms into larger ones. We believe that observing a larger MM seasonal cap might be a leading indicator for this type of storm merger and eventual regional dust storm (that can combine

with other latitudes to form a GDS). One possibility is that before the onset of the GDS, the southern polar region may be colder than usual, allowing for the MM seasonal cap to remain larger than the seasonal average. This could lead to a larger frost outlier, which may sublime faster. The rapid sublimation of the larger frost outlier might result in more intense storms or a higher frequency of storms, potentially triggering the MY 28 GDS event. Further analysis is required to thoroughly test this hypothesis.

3.5 Conclusion/Summary

We utilized daily MARCI mosaics in conjunction with an open-source computer vision library to track the SPSC for MYs 29-31, spanning from $L_s = 175^\circ$ to $L_s = 340^\circ$. Employing the HSV colour format seen in Chapter 2, our method dynamically adjusts colour bounds to enhance the tracking of the SPSC during significant regional dust storms. We applied an ellipse fit to determine the average latitude of the SPSC and analyzed the seasonal variability of this latitude using both a line of best fit and a non-parametric smoothing function, achieving a temporal resolution of approximately 1° of L_s date. Additionally, to examine the latitudinal recession rate, we calculated the first derivative of the SPSC latitude extent to identify periods of acceleration and deceleration.

Although our approach is slower than the static method seen in Chapter 2, it is considerably more efficient than previous methods that manually outlined the SPSC. Our technique faithfully replicates prior measurements but tends to yield slightly larger SPSC areas by approximately 0.3 latitudes. Over 95% of total MARCI mosaics were utilized, discarding only those with irreparable artifacts and poor ellipse fit. Additionally, Calvin et al. (2017) had outlined the SPSC for the same MYs; however, they did not determine the seasonal recession rates for those MYs.

We segmented the recession into two phases based on the $L_s = 260^\circ$ discontinuity, marking the separation of the MM seasonal cap from the main SPSC. Pre-discontinuity, most MYs exhibit consistent behaviour, divided into five distinct periods: $L_s = 175.3^\circ - 194.6^\circ$, $L_s = 194.6^\circ - 215.4^\circ$, $L_s = 215.4^\circ - 234.5^\circ$, $L_s = 234.4^\circ - 254.0^\circ$, and $L_s = 254.0^\circ - 260^\circ$, delineated by respective retreat

rates. The fastest annual latitudinal recession typically occurs around $L_s = 235^\circ$, with the second fastest around $L_s = 252^\circ$. Post-discontinuity, MYs show more variability, with the SPSC receding linearly until $L_s = 307^\circ$ before gradually nearing the residual cap latitude for the remaining period. Another minor discontinuity associated with the late-summer outlier shows significant variability in timing and magnitude compared to the MM discontinuity.

A critical observation from MY 28 revealed significant deviations in the recession rate—markedly higher—and size—substantially smaller—of the SPSC, coinciding with the MY 28 GDS event. This suggests a warmer atmosphere during the storm amplifies the sublimation and retreat rate, resulting in a diminished cap subsequently. Additionally, the MM seasonal cap outlier in MY 28 began significantly larger than the interannual average, later undergoing faster sublimation and concluding earlier than typical. This observation is the first time the outlier is seen bigger before the onset of the MY 28 GDS event.

3.6 Conclusion/Summary

Chapter 4

The formation and evolution of the Cool and Bright Anomalies and Warm and Dark Anomalies on the North Polar Residual Cap

4.1 Background Information and Motivation

4.1.1 Background Information/Motivation

As observed by Kieffer and Titus (2001), thermal observations were employed to monitor the north polar region of Mars. From these observations during MY 28, a notably cold and bright site located at 87° N, 313° E was identified and subsequently termed the Cold and Bright Anomaly (CABA). At this site, during early summer at a L_s of 104° , temperatures were observed to be approximately 10 K cooler than the surrounding areas. Kieffer and Titus (2001) hypothesized that this cooling effect results from "cold trapping," where small, highly reflective ice grains condense on the surface, effectively reflecting sunlight and thereby slowing the temperature rise. This mechanism not only maintains the surface cooler but also preserves its high reflectivity (albedo), thus sustaining a positive feedback loop. By late summer, around $L_s = 147^{\circ}$, the site's albedo diminished by 30%, from 0.5 to 0.35, compared to a surrounding albedo of 0.4. During this period,

Kieffer and Titus (2001) noted three significant darkening events between $L_s = 140^\circ$ and $L_s = 160^\circ$, with the site transitioning to slightly warmer temperatures than its surroundings. These sites are referred to as Warm and Dark Anomaly (WADA), as the site becomes both warmer and darker compared to their surroundings.

Cantor et al. (2010), utilizing wide-angle imagers such as the Mars Orbital Camera and MARCI, identified the same CABA site, which they named the Vostok site, along with several bright spots across the NPRC during early summer for MYs 28-29. In the late summer transformation period, Cantor et al. (2010) also observed the same three darkening events and spectral observations indicated that the WADA site exhibited strong water-ice signatures.

Neither study provided details on the exact formation of the localized cold spots required for the cold trapping process nor the transformation from CABAs into WADAs. Moreover, while Cantor et al. (2010) acknowledged the existence of other potential CABA sites, these were not extensively studied. The primary aim of this chapter is to propose a novel mechanism for the formation and evolution of CABAs and WADAs. Additionally, this chapter seeks to characterize the evolution of multiple CABA and WADA sites through both thermal and visual observations

4.2 Methods

4.2.1 MARCI Observations

This chapter uses the same MARCI mosaics discussed in Chapter 2.1. The MARCI mosaics enable me to characterize the albedo evolution of many sites across the NPRC. Previous studies had focused only on the Vostok CABA sites; however, MARCI reveals many CABA sites across the residual cap (Figure 4.1). Multiyear analysis shows many recurring CABA and WADA sites; however, for this dissertation, we identify five such sites (Figure 4.1; Table 4.1). We have identified five non-anomalous sites, each located within 60 kilometers of the corresponding anomaly sites, that do not display any behaviours characteristic of CABAs. Additionally, we have pinpointed five sites that exhibit high levels of surface activity, such as dust streaks, indicative of heightened

surface wind activity (Figure 4.1; Table 4.1). The non-anomaly sites and high dust streak sites are control sites, where the seasonal behaviour is relatively well understood compared to the anomaly sites. For all sites, we characterize the surface albedo by measuring the average pixel value within a 2x2 box centered at each site. This method employs a 2x2 box rather than a single pixel to mitigate the effects of transient, short-lived storm events. By averaging over a slightly larger area, we buffer against the disproportionate impact of these storms on the albedo measurements. The average pixel value is calculated from the Red, Green, and Blue pixel values captured in the mosaics and stored as PNG files. Each colour channel's values are encoded as an 8-bit integer, ranging from 0, indicating no signal, to 255, representing the maximum signal. This average pixel value serves as an indicator of albedo changes, although it is subject to influence by atmospheric conditions, as depicted in Figure 4.2.

Table 4.1: List of longitude and latitude of all of the locations for the CABA and WADA sites, non-anomaly sites, and high dust streak sites.

Type	Longitude [° E], Latitude [° N]	Site Comment
CABA and WADA #1	222.1, 85.1	Vostok anomaly site
CABA and WADA #2	312.2, 86.7	
CABA and WADA #3	307.2, 85.7	
CABA and WADA #4	306.1, 84.1	
CABA and WADA #5	350.7, 85.2	
Non-anomalous #1	211.5, 86.0	Near CABA and WADA #1
Non-anomalous #2	296.4, 86.9	Near CABA and WADA #2
Non-anomalous #3	300.1, 85.2	Near CABA and WADA #3
Non-anomalous #4	304.0, 84.4	Near CABA and WADA #4
Non-anomalous #5	359.4, 85.5	Near CABA and WADA #5
High Dust Streak #1	94.6, 86.5	
High Dust Streak #2	182.2, 86.7	
High Dust Streak #3	345.5, 86.6	
High Dust Streak #4	130.6, 86.6	
High Dust Streak #5	137.0, 86.1	

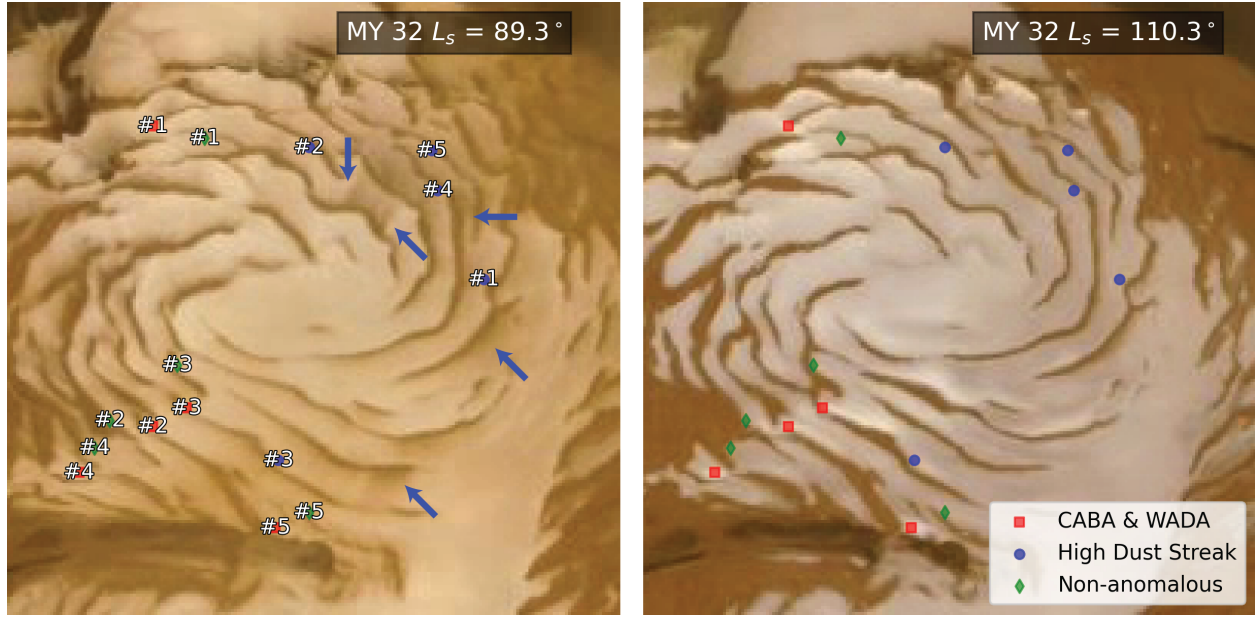


Figure 4.1: MARCI mosaic of a portion of the NPRC during MY 32: The left image depicts the NPRC in late spring, while the right image shows it in early summer. The right image contains many bright spots associated with CABAs; however, this chapter focuses only on five CABA and WADA sites marked with red squares. Nearby non-anomalous sites are indicated by green rhomboids, and sites with high dust streaks are marked by blue circles. In panel A, the numbering of sites #1 to #5 is shown, and blue arrows highlight typical dust streaks that frequently occur during late spring ($L_s = 80^\circ$ to 90°).

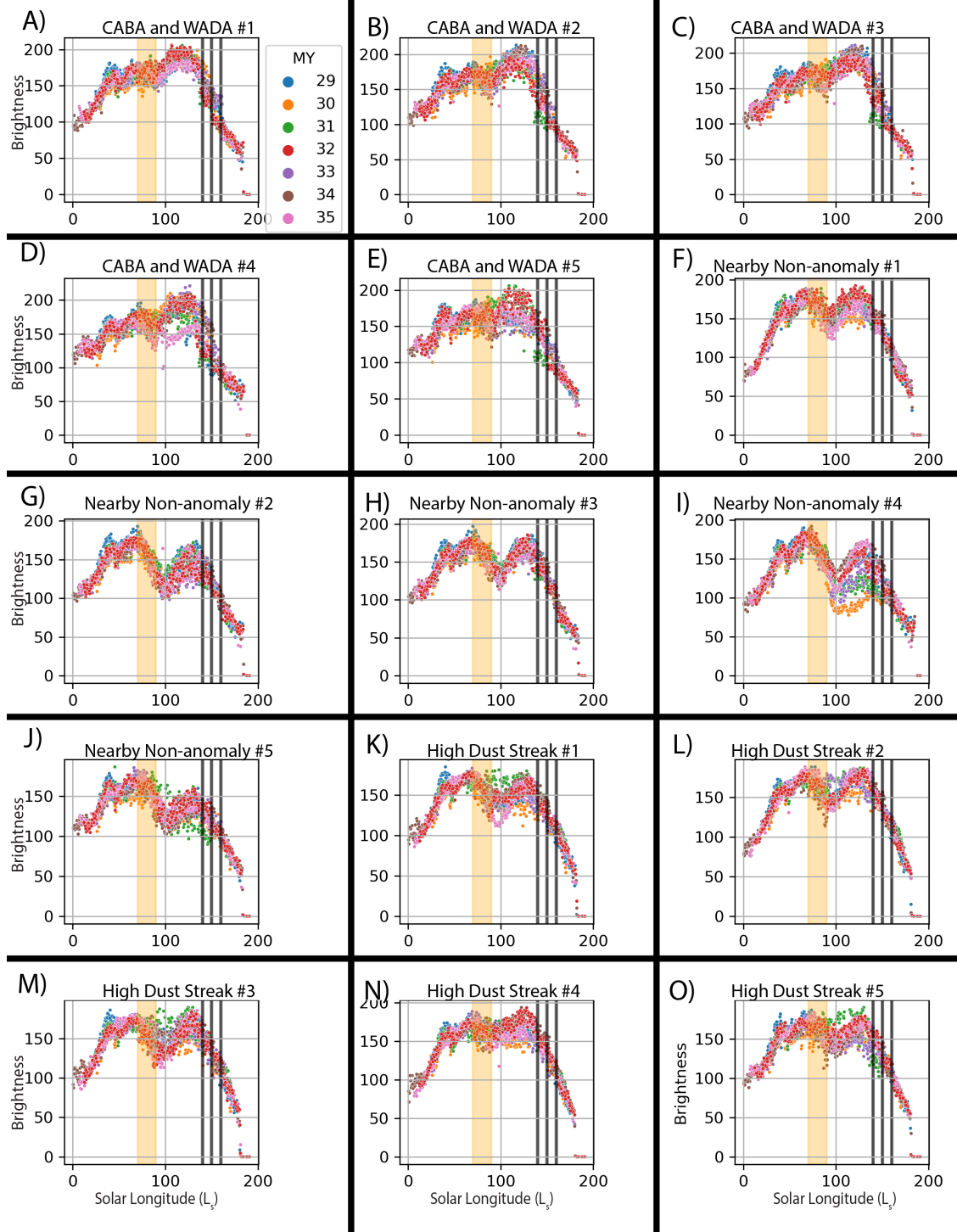


Figure 4.2: Mean brightness evolution between the CABA and WADA sites and the corresponding non-anomalous sites and high dust streak sites, as listed in Table 4.1. The x-axis represents the L_s [°], temporal resolution of 0.5° of L_s dates, and the y-axis represents the pixel brightness recorded by the MARCI mosaics, saved as PNGs, at the sites. The orange-shaded area indicates the period of peak katabatic wind activity while the three black lines indicate the major darkening events. Between $L_s = 90^\circ$ and $L_s = 108^\circ$, the mean brightness decreases at all non-anomaly sites and high dust streak sites (I-O), while it remains constant at the anomaly sites (A-E). Short periods of decreases are due to short-lived storms passing over the sites, resulting in a decrease in the mean brightness values.

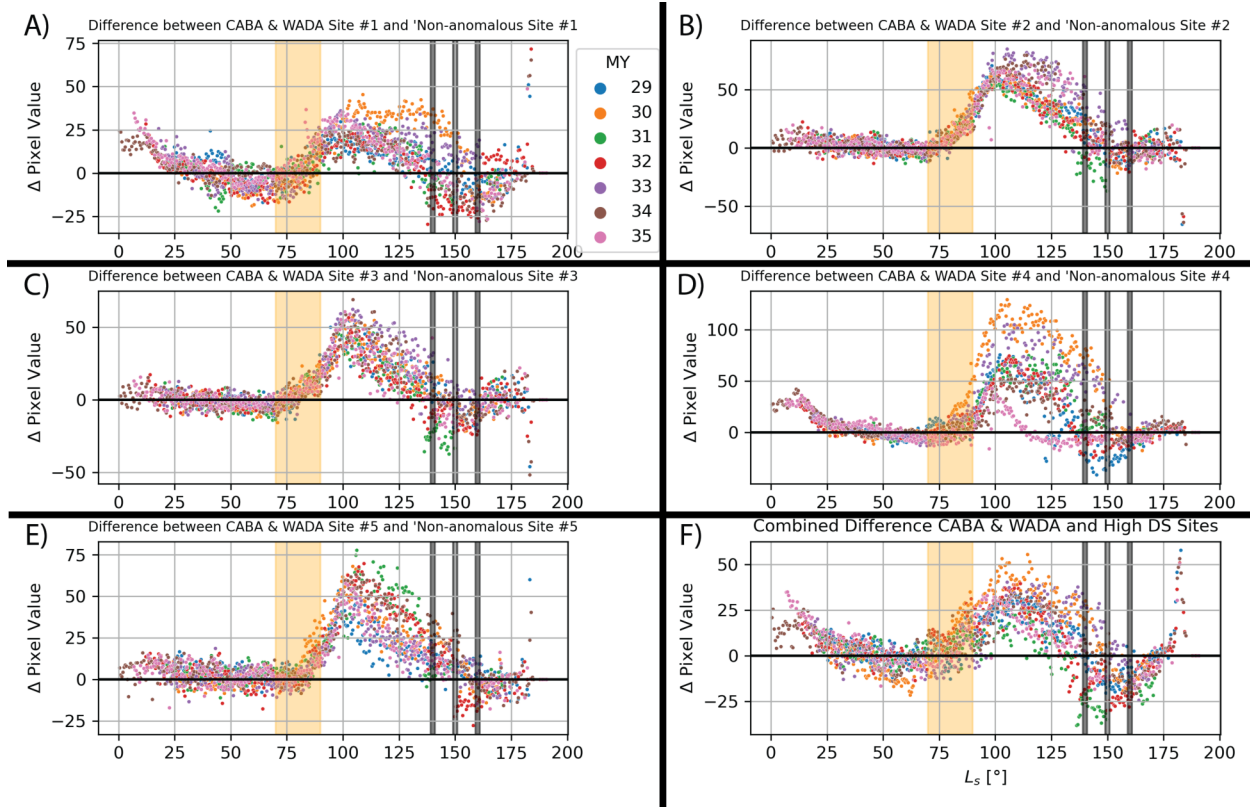


Figure 4.3: Mean brightness difference between the CABA and WADA sites and their corresponding nearby non-anomalous sites, as listed in Table 4.1. The x-axis represents the L_s [°] and the y-axis represents the difference (Δ) in the pixel value. The orange-shaded area indicates the period of peak katabatic wind activity while the black lines indicate the three major darkening events. Positive values represent the anomaly sites being brighter than the non-anomaly sites, whereas negative values represent the opposite. The difference begins to increase during the peak katabatic wind season $L_s = 75^\circ$. Coinciding with the darkening events, there are sharp decreases in difference values around the time of the darkening events.

4.2.2 THEMIS Observations

The Thermal Emission Imaging System (THEMIS) aboard the 2001 Mars Odyssey has been providing detailed observations of Mars' polar regions since MY 25 (Christensen et al. 2001). Operating in both infrared (IR) and visible (VIS) modes, THEMIS plays a crucial role in acquiring precise spatial data by measuring the emittance, the light that is emitted by the Martian surface. In IR mode, it captures wavelengths from 6.78 microns to 14.88 microns with a resolution of 100 meters per pixel, enabling more detailed analysis of anomaly sites and surpassing previous studies' capabilities (Christensen et al. 2001). The VIS mode records wavelengths from 0.425 microns to

0.860 microns at 18 meters per pixel, offering finer details than the Mars Color Imager (MARCI) at about three kilometers per pixel, albeit at a lower temporal resolution.

THEMIS VIS images at the sites of interest are essential in distinguishing between the anomaly and non-anomalous sites. The period of interest is before and after the emergence of CABAs and WADAs, providing insights into potential processes behind these anomalies (Figure 4.4). The higher resolution of the VIS images provides a finer view of the surface and surface textures. In addition, we use THEMIS VIS images from MYs 32-35 spanning $L_s = 30^\circ$ to $L_s = 120^\circ$ to categorize surface clouds observations, similar to work from by Smith et al. (2013). These observations are combined with Smith et al. (2013) observations, which focused on MYs 29-31, to generate a map of surface elongated cloud activity during northern spring (Figure 4.5). As mentioned in Section 1.2.2, these clouds are associated with katabatic winds and serve as an indicator for near-surface wind activity - a higher number of clouds correlate faster katabatic winds (Smith et al. 2013; 2015). We utilize THEMIS IR observations, specifically from $L_s = 30^\circ$ to $L_s = 170^\circ$, to monitor surface temperatures at various sites of interest (Figure 4.6). Surface temperature measurements will be able to recreate observations from Kieffer and Titus (2001) and provide insights into potential processes behind these anomalies. We extract the surface temperature at the sites using the JMars software, which includes the functionality of extracting surface temperature measured by the IR observations (Christensen et al. 2009). It is important to note that in the late summer period, post- $L_s = 130^\circ$, the observational dataset becomes sparse, with significant gaps. The quantity of observations varies considerably, ranging from as few as six data points at some sites during MY 30, to a maximum of twelve points during MY 32 (Figure 4.7).

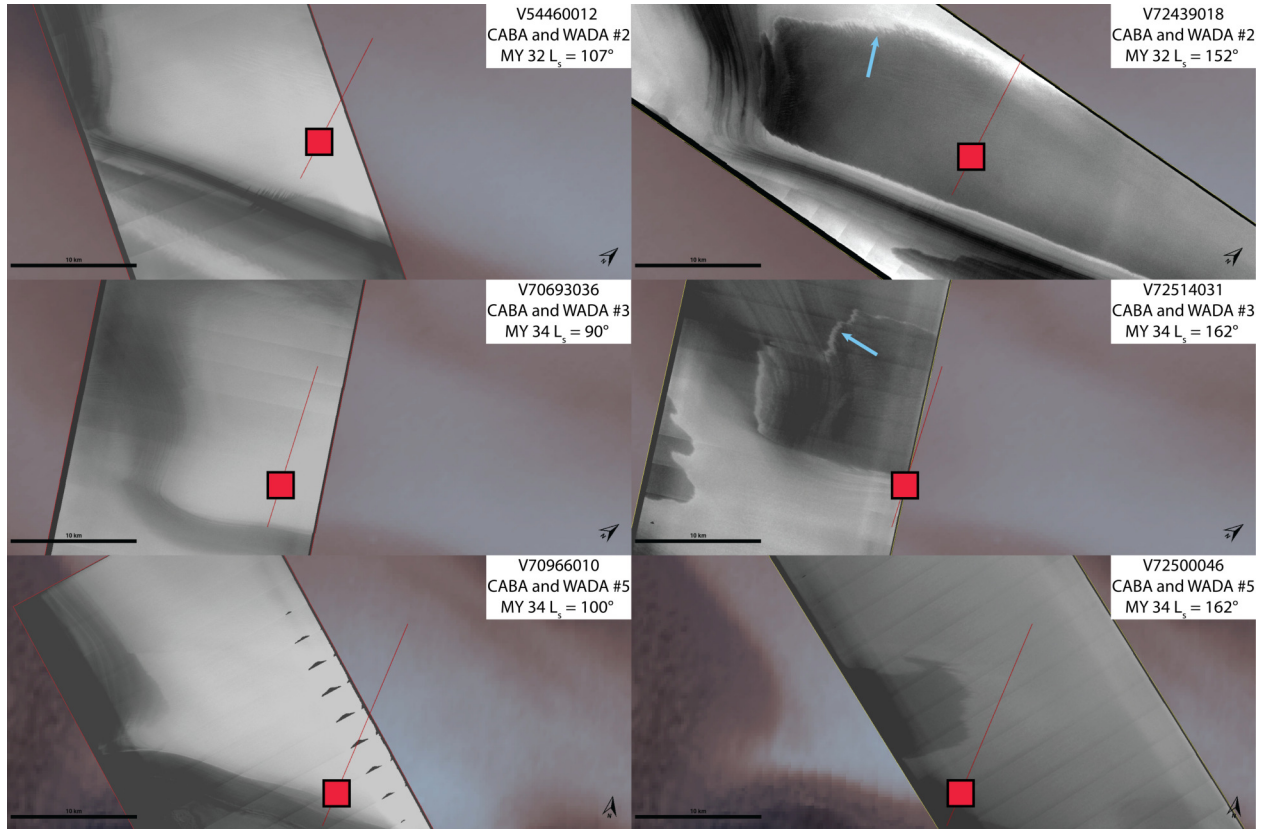


Figure 4.4: Composite of visual THEMIS images in band #3, centered at 650 nanometers, superimposed on the Viking Colorized Global Mosaic, arranged in a 3x2 grid showcasing the progression of various CABA and WADA sites from early summer (Left Column), during the time of the CABAs, and late summer (Right Column), during the time of the WADAs. The red square in each sub-figure presents the site location. The red lines across each image trace the topography profiles shown in Figure 4.9, running perpendicular to the nearby troughs. Light blue arrows in each sub-figure highlight the bright halos that develop around WADAs during this transition period.

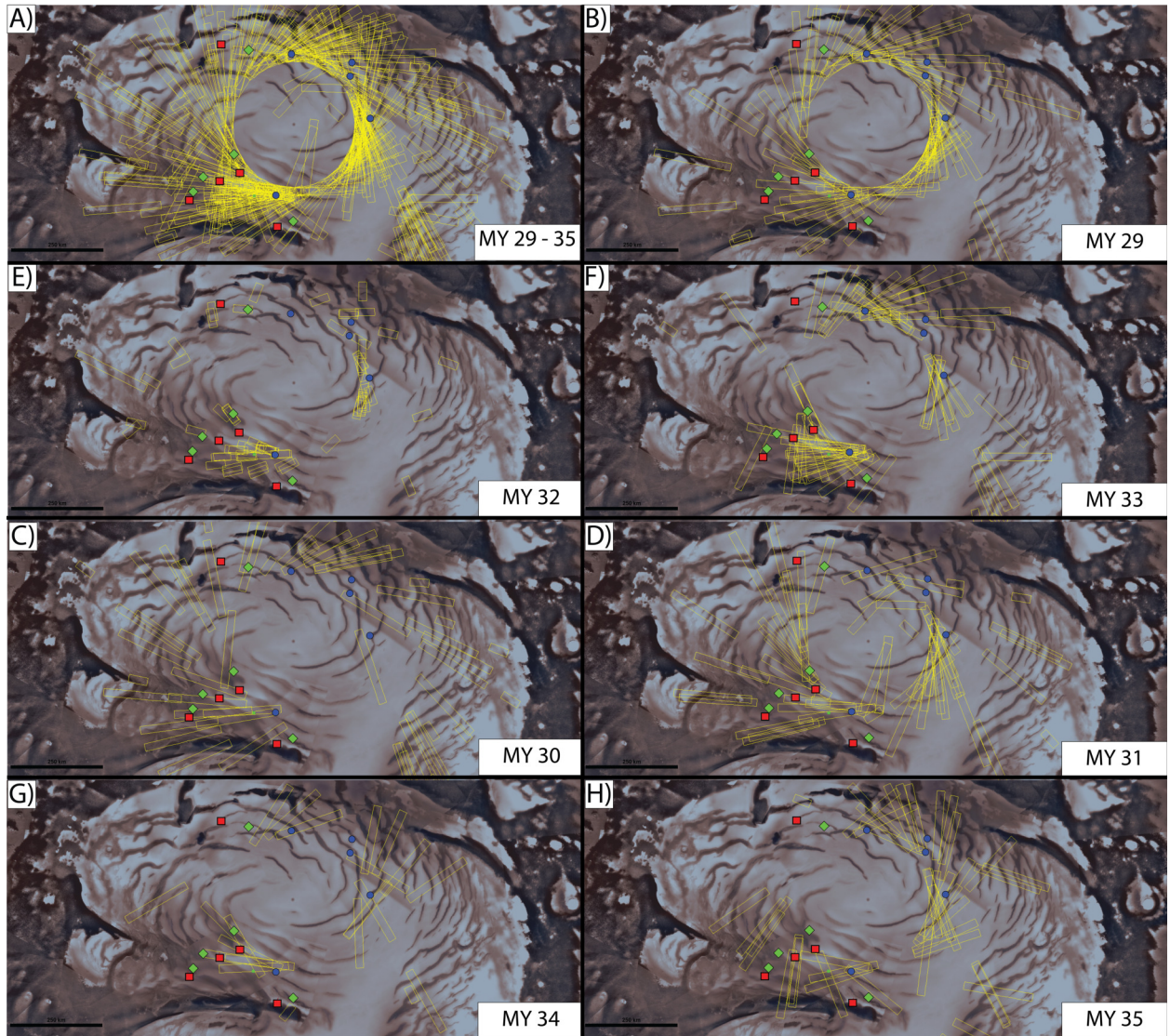
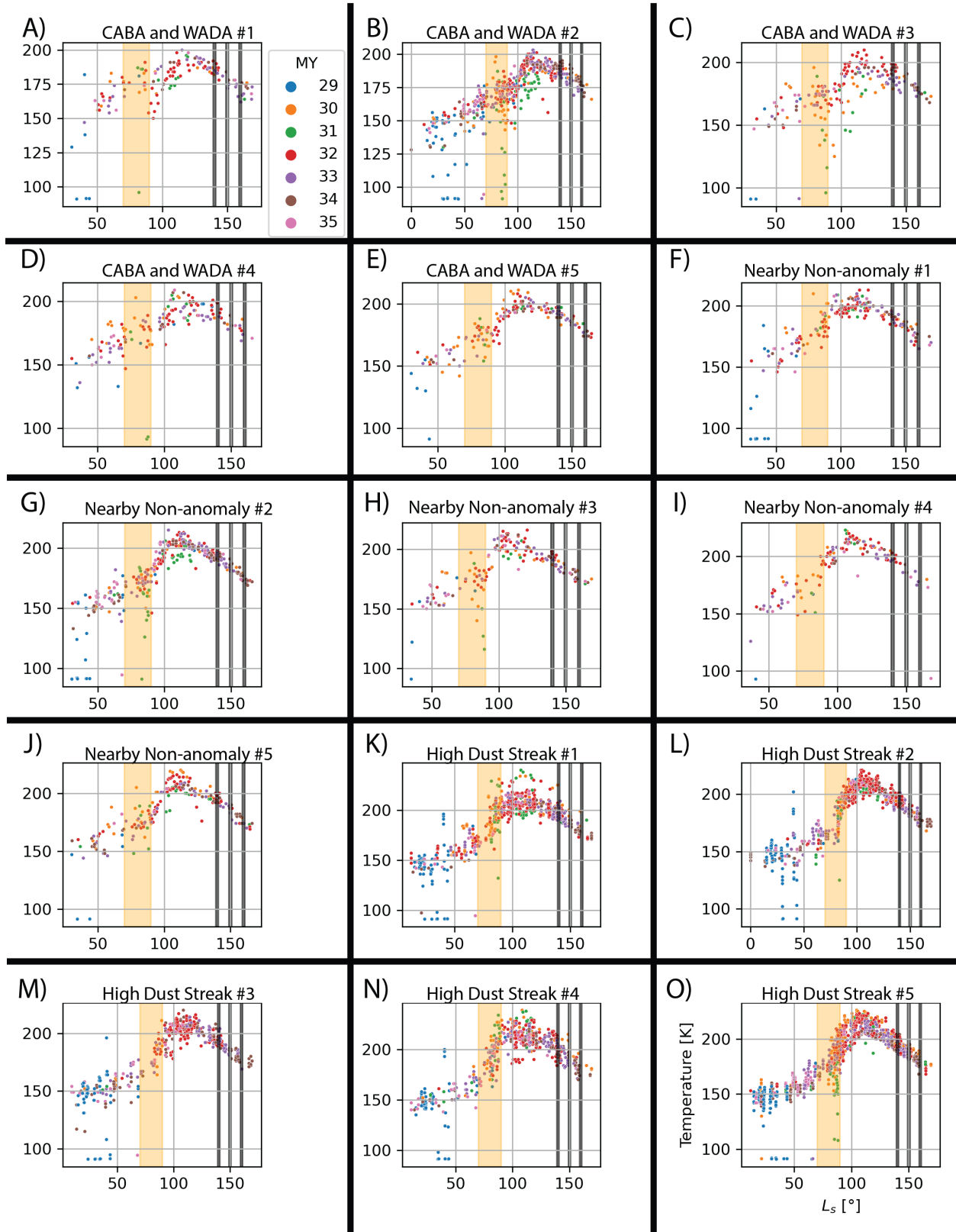


Figure 4.5: Viking Colorized Global Mosaics with overlaid THEMIS image outlines (indicated by yellow outlines) representing surface cloud observations. These sub-figures illustrate the distribution of clouds across MYs 29-35. In each sub-figure, the locations of interest are marked with coloured shapes: red squares for CABA and WADA sites, green rhomboids for nearby non-anomalous sites, and blue circles for high dust streak sites. The yellow outlines show that clouds, while present across various sites, are comparatively less common at the anomaly sites marked with red squares and most prevalent at the high dust streak sites marked with blue circles.



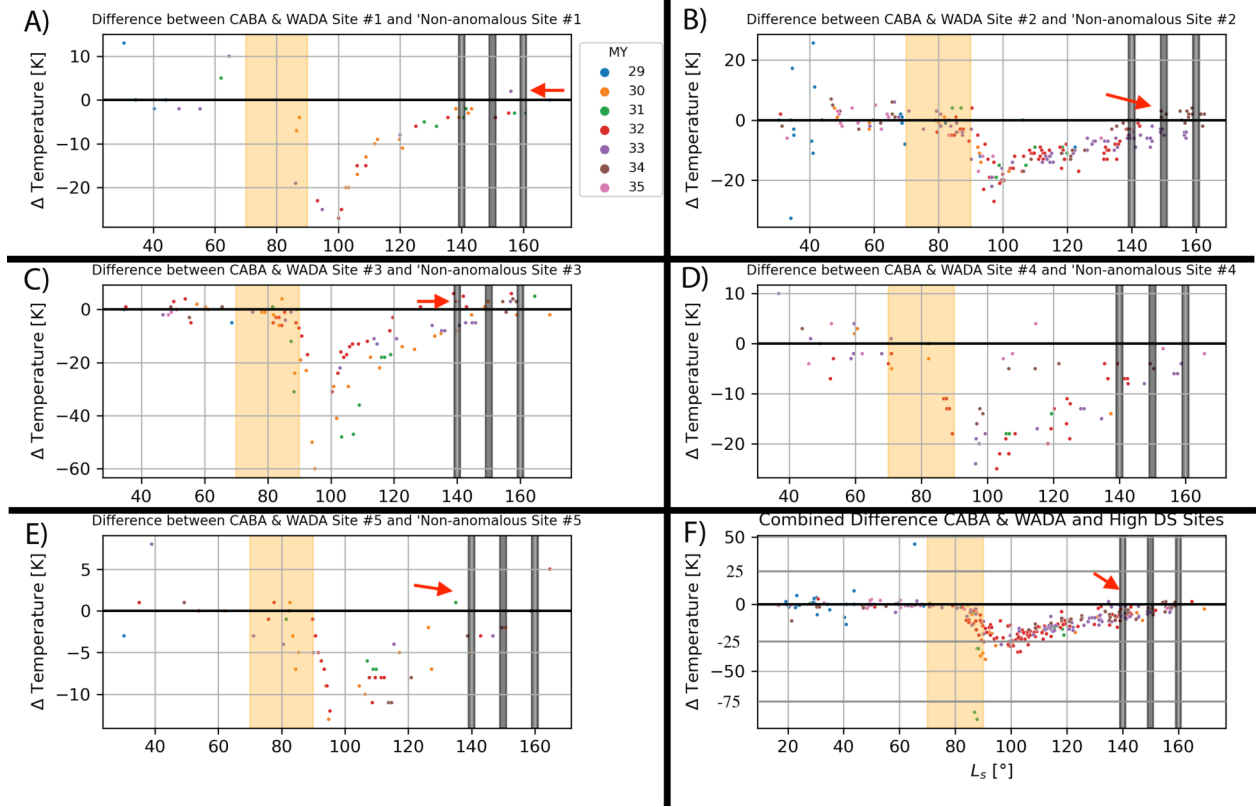


Figure 4.7: Temperature differences between anomalous sites and their corresponding non-anomalous and high dust streak sites. Negative values on the graph indicate cooler temperatures at the anomaly sites compared to others. The x-axis represents the L_s [°] and the y-axis represents the surface temperature [K]. The orange-shaded area highlights the peak katabatic wind season and the three dark lines signify the major darkening events. A notable divergence in temperature between anomaly and non-anomaly sites begins around $L_s = 82^\circ$, with the largest annual temperature difference reaching approximately 15 K by $L_s = 105^\circ$. During the period of WADAs, the anomaly sites are warmer than the non-anomaly sites, ranging from 1-8 K with an average of 5 K (Red Arrows).

4.2.3 HiRISE and CTX Observations

The High Resolution Imaging Science Experiment (HiRISE) and the Context Camera (CTX), aboard the Mars Reconnaissance Orbiter, provide images with resolutions of approximately 30 centimetres per pixel and six meters per pixel, respectively (Bell et al. 2013; McEwen et al. 2007). These resolutions notably surpass those of MARCI and THEMIS, offering enhanced observations of the surface as anomalies emerge (Figure 4.8). However, both instruments have inferior temporal resolutions compared to MARCI and THEMIS, with observations limited after $L_s = 130^\circ$. These gaps significantly challenge our ability to perform comparisons of the anomaly sites before and after

the anomalies appear. To mitigate these limitations, we incorporate additional CTX and HiRISE observations from MY 36, thereby facilitating comparative analysis of the anomaly sites.

4.2.4 MOLA Topography

Additionally, the Mars Orbiter Laser Altimeter (MOLA) on the Mars Global Surveyor provides high-resolution topographical maps of Mars, achieving a vertical resolution of 1.5 meters (Zuber et al. 1992). We focus on the local topography surrounding the anomalous and non-anomalous sites. Specifically, we analyze topographic profiles that extend 8 to 12 kilometers, centred on these sites and oriented perpendicular to the nearby troughs (Figures 4.8, 4.9). For extracting this topographical data, we utilize the JMars software and its built-in functionality for precise topographical profiling (Christensen et al. 2009).

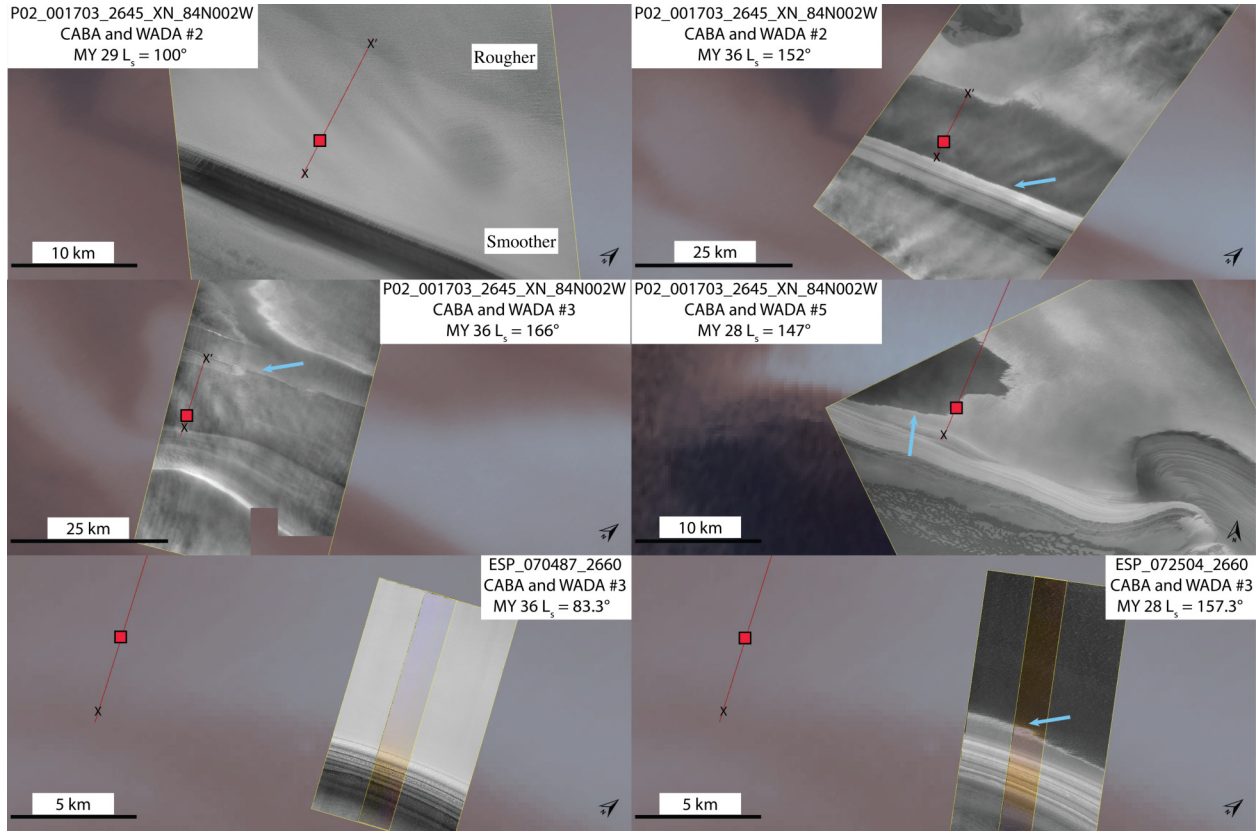


Figure 4.8: CTX (Top two rows) and HiRISE (Bottom row) observations showcasing the transformation of CABA and WADA sites, highlighted by blue arrows indicating the bright halos at WADA sites. The top row displays CABA and WADA site #2 during the peak surface temperature difference (Left) and the second darkening event (Right), illustrating the site's smoother surface and fewer streaks compared to its surroundings during summer. The middle row captures CABA and WADA sites #3 and #5 as the transition of the site into WADAs, including a dust storm observed at site #3 post-final darkening event, surrounded by bright halos. The bottom row focuses on the CABA and WADA site #3 before and after its transformation into a WADA, noting a rougher surface post-transition and halo streaks spanning about 650 meters.

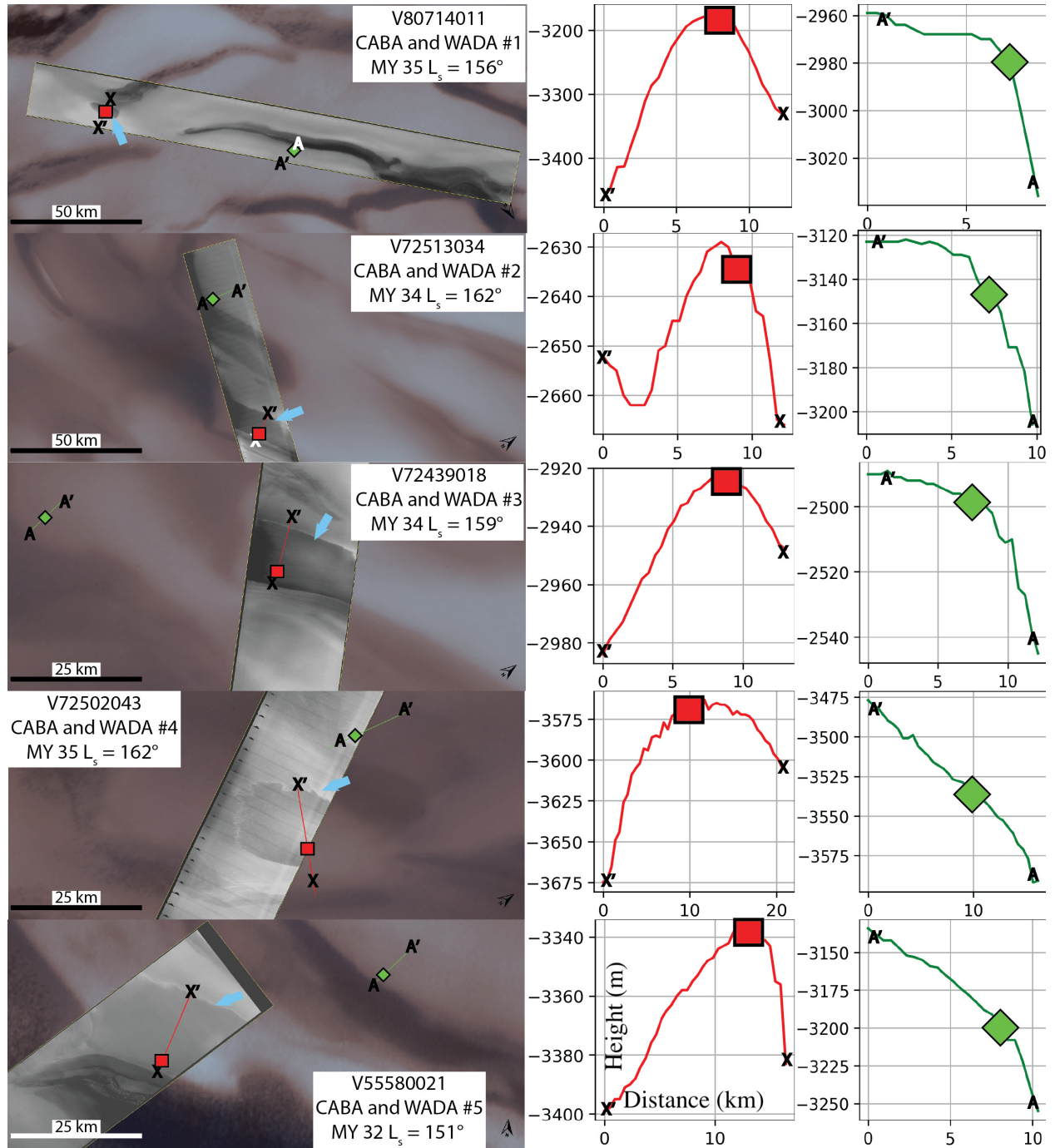


Figure 4.9: THEMIS images of WADAs overlaid on the Viking colour mosaic, with red lines illustrating the topographic cross-sections at the CABA and WADA sites (the x-axis represents the distance [km] and the y-axis represents the elevation [m]), all of which are located on local rises ranging from 30 to 100 meters. The highest elevation is at CABA and WADA site #1, and the smallest is at site #2. In contrast, the green lines represent the cross-sections of non-anomalous sites, which are not situated on similar topographic elevations, highlighting a distinct topography at the CABA and WADA sites.

4.2.5 Mesoscale Simulations

The Mars Mesoscale Model (MMM), developed by the Laboratoire de Météorologie Dynamique (LMD), serves as a cornerstone for our analysis of Martian atmospheric dynamics, particularly focusing on near-surface wind phenomena. This model's architecture is built upon the Advanced Research Weather Research and Forecasting (ARW-WRF) model, which is notably adapted to accommodate the distinct environmental conditions of Mars (Spiga and Forget 2009).

4.2.5.1 Physical Assumptions and Capabilities

MMM is tailored specifically to Mars by incorporating non-hydrostatic equations, allowing it to simulate atmospheric motions at scales where hydrostatic assumptions typically applied in larger-scale models (like global climate models) break down. This feature is crucial for accurately resolving small-scale dynamical processes that occur in the Martian atmosphere, including the development of katabatic winds and local dust devils, which are influenced by the planet's complex topography and thin atmosphere (Spiga and Forget 2009).

The model employs a detailed representation of Martian atmospheric physics, which includes:

- **CO₂ Gas Infrared Absorption/Emission:** The radiative transfer modelling, including CO₂ absorption and emission, is based on the parameterization by Hourdin (1992) and Forget (1998) for infrared dust absorption, emission, and diffusion. This is crucial for understanding Martian atmospheric dynamics due to the predominant CO₂ composition.
- **Thermal Inertia and Soil Model:** The simulation of soil thermal conduction is implemented using an 11-layer soil model developed for Mars by Hourdin et al. (1993). This model considers soil density and soil-specific heat capacity as constants, which are essential for accurately simulating the Martian surface temperature dynamics.
- **Dust Lifting and Sedimentation:** Dust cycle modelling in the MMM includes dust lifting and sedimentation processes based on the works by Forget et al. (1999) and Newman et al.

(2002). These studies provide a basis for understanding how dust interacts with the Martian atmosphere and influences various climate and visibility factors.

- **Water Ice Clouds and Sublimation/Deposition:** The parameterization of the water ice cycle, particularly the processes of sublimation and deposition, is based on Montmessin et al. (2004). This work details how water ice clouds form and behave in the Martian atmosphere, which is vital for modelling local and regional weather patterns on Mars.

These parameterizations are integrated with a dynamical solver that is robust enough to handle Mars's lower gravity and lack of a thick atmosphere, enabling the model to simulate the unique wind patterns observed on Mars. For example, the model effectively captures intense nighttime katabatic flows that develop along crater walls and steep topographies such as the spiral troughs, as well as predicting the clouds associated with these winds (Smith et al. 2013; 2015; Smith and Spiga 2018; Spiga and Smith 2018).

4.2.5.2 Simulation Parameters and Operational Constraints

Our simulations, covering Martian solar longitude $L_s = 30^\circ$ to $L_s = 165^\circ$, are designed to understand seasonal variations in atmospheric dynamics. Each simulation run spans two Martian days, simulating every 5° of L_s , with a focus on recording meridional and zonal winds after a 24-hour stabilization period to ensure the atmosphere in the model reaches a quasi-steady state. This stabilization is critical, given the sensitivity of the Martian atmosphere to initial conditions due to its thin composition and the strong solar heating effects at the surface (Smith et al. 2015; Spiga and Smith 2018).

From $L_s = 30^\circ$ to $L_s = 140^\circ$, we use the same simulations as those utilized by Smith and Spiga (2018), featuring a grid size of 18 kilometers and a timestep of 40 seconds, across the same total domain of 3200x3200 kilometers (Figure 4.11). We have built upon Smith and Spiga (2018) simulation runs by including $L_s = 145^\circ$ to $L_s = 165^\circ$. The new simulations maintain the same domain parameters but with a reduced timestep of ten seconds. This adjustment is crucial for the stability of the model during late summer. Our analysis reveals that wind speeds during this

period are sufficiently high to violate the Courant-Friedrichs-Lewy (CFL) stability criterion for the employed time-marching simulations such as atmosphere models. The CFL condition dictates that the chosen time step must be proportionally smaller than the fixed grid size to ensure numerical stability (Courant et al. 1967).

Despite the model's capabilities, it is important to acknowledge the limitations inherent to the current version of the MMM, particularly the absence of a nesting feature, which would allow for finer spatial resolution. Additionally, stability challenges persist under certain configurations and parameter combinations. Nevertheless, within the constraints of a ten-second timestep and an 18-kilometer grid size, our simulations effectively depict the large-scale transient eddies crucial to our late-summer analyses (Figure 4.11).

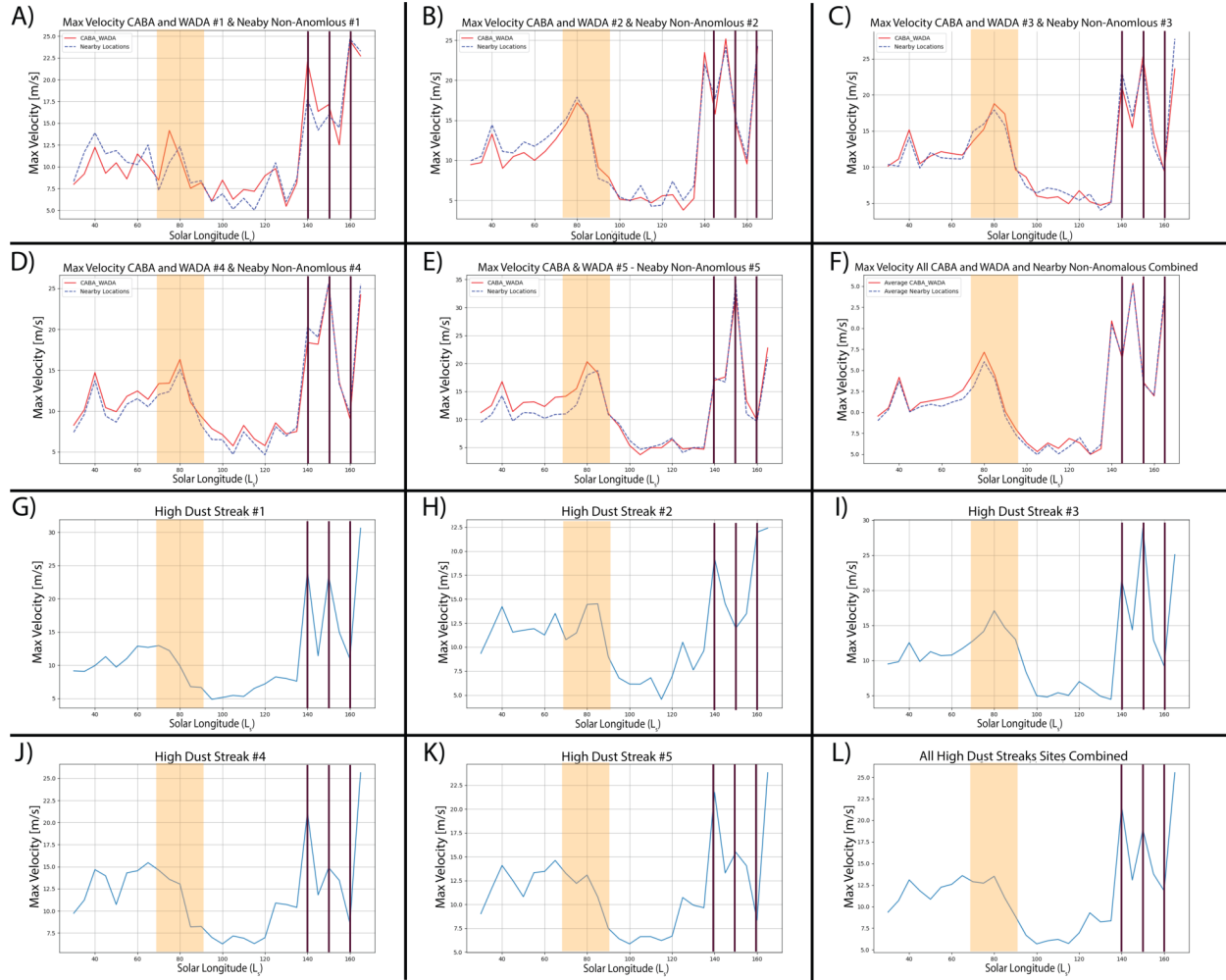


Figure 4.10: MMM surface wind outputs at all sites shown in Table 4.1. The x-axis represents the L_s [°] and the y-axis represents the max velocity in the simulated L_s date. The orange-shaded area represents the peak katabatic wind season and three dark lines mark the major darkening events. The first and second rows show the anomaly sites (red) and non-anomaly sites (green) on the same plot. During late summer, winds can exceed 20 meters per second, each peak winds occurring during the darkening events.

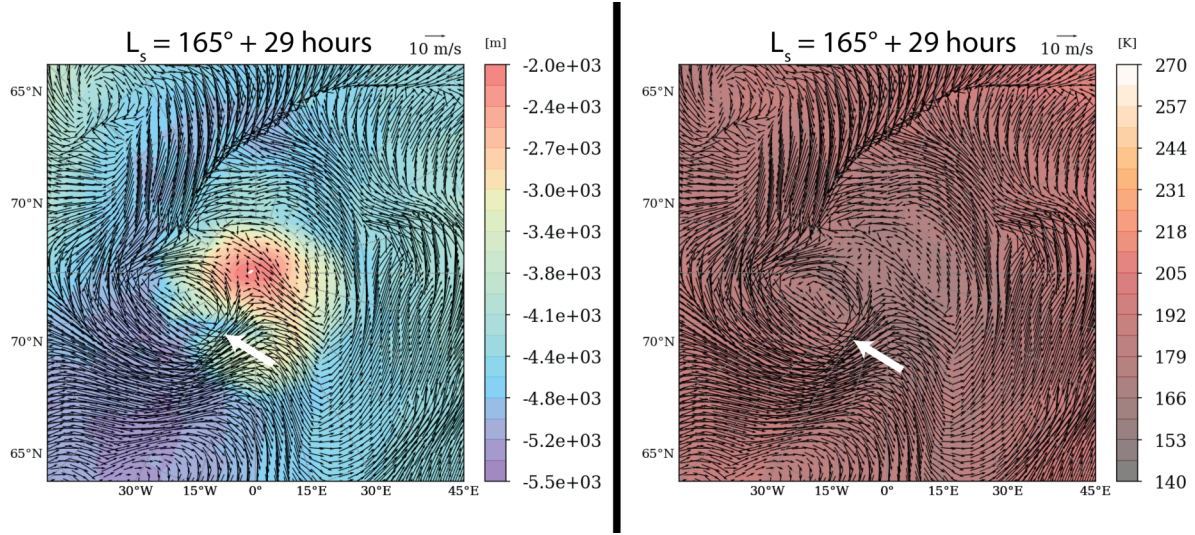


Figure 4.11: MMM domain simulated domain during $L_s = 165^\circ + 4$ hours. The left shows the wind field overlaid on top of the topography and the right shows the wind field on top of the surface temperature. Large eddies (white arrows mark an example) and strong winds are seen during the period after $L_s = 140^\circ$ where the winds exceed 20 meters per second

4.3 Results

4.3.1 Evolution Observed by MARCI

Before $L_s = 75^\circ$, our analysis indicates that all sites display largely uniform characteristics, exhibiting only minimal variations in brightness levels (Figures 4.2, 4.3). A distinct divergence in brightness between the anomaly sites and their non-anomalous counterparts emerges around $L_s = 75^\circ$, with the anomaly sites consistently exhibiting higher brightness levels. The divergence date varies slightly, occurring as early as $L_s = 72^\circ$ and as late as $L_s = 78^\circ$. The brightness difference reaches an annual maximum at approximately $L_s = 108^\circ$, coinciding with the emergence of numerous bright sites across the NPRC (Figures 4.3, 4.12, 4.13). After the annual maximum, the brightness difference gradually diminishes until $L_s = 140^\circ$. Notably, we document three major darkening events at $L_s = 140^\circ$, $L_s = 150^\circ$, and $L_s = 160^\circ$, along with several minor events (Figures 4.3, 4.12, 4.13). The major events consistently occur around similar L_s dates, exhibiting a variability of $\pm 5^\circ$ L_s dates. During these events, the brightness at the anomaly sites decreases by 38%, consistent with

prior observations at the Vostok anomaly site. Notably, MY 34 does not witness the final darkening event. A similar trend is observed in the brightness difference between the average anomaly site and high dust streak sites, with a divergence date around $L_s = 70^\circ$, and an annual maximum difference at $L_s = 110^\circ$. It should be noted that during a few MYs, some of the non-anomalous sites transition into WADA sites after $L_s = 165^\circ$.

In the mosaic, most CABAs are consistently on the higher side of nearby troughs (the side closest to the north pole) (Figures 4.1, 4.12, 4.13), especially the more pronounced ones between longitudes 305° E and 320° E (Figures 4.12, 4.13). There is significant annual variation in both the number of CABAs and their contrast against the surrounding terrain. MY 30 exhibits the highest count of CABA sites and the most distinct contrast, while MYs 31 and 32 record the fewest. During the darkening events, all CABA sites transform into WADAs, although some WADA sites do not display any CABA-like behaviour before this transformation. WADAs are generally located between longitudes 230° E and 340° E, exhibiting more annual variability compared to CABAs. The spatial extent of WADAs varies widely, from covering about half of the NPRC to confined sections. Intriguingly, each darkening event is followed by a significant increase in both the number and extent of WADAs. Interestingly, the newly formed WADAs, especially during major darkening events, frequently coincide with nearby storms that intensify with each event. MYs 31 and 32 show the most extensive coverage of WADAs, whereas MY 34 shows the least (Figures 4.12, 4.13).

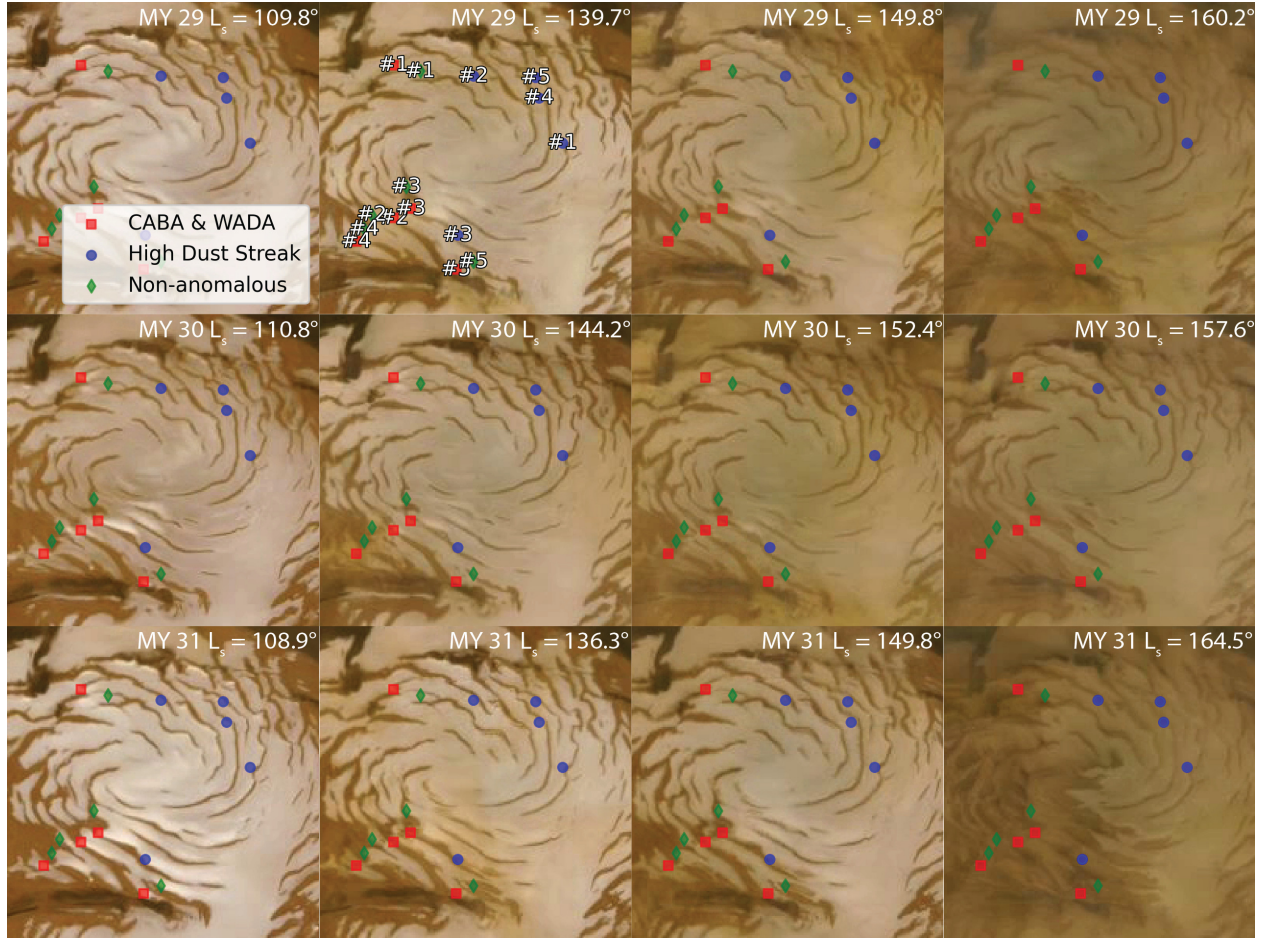


Figure 4.12: MARCI mosaic crop for MYs 29-31 showing key moments in the evolution of the anomalies. The first column shows the portion of the NRPC, centered at 90° N, during early summer when the CABAs appear throughout the region. The second column shows the same portion but during the first darkening event with some CABAs undergoing the WADA transformation. The third column shows the second darkening event with the number and size of WADAs increasing compared to the previous event. The fourth column shows the final darkening event, the largest darkening event where the WADAs can cover about half of the NRPC as seen during MY 31.

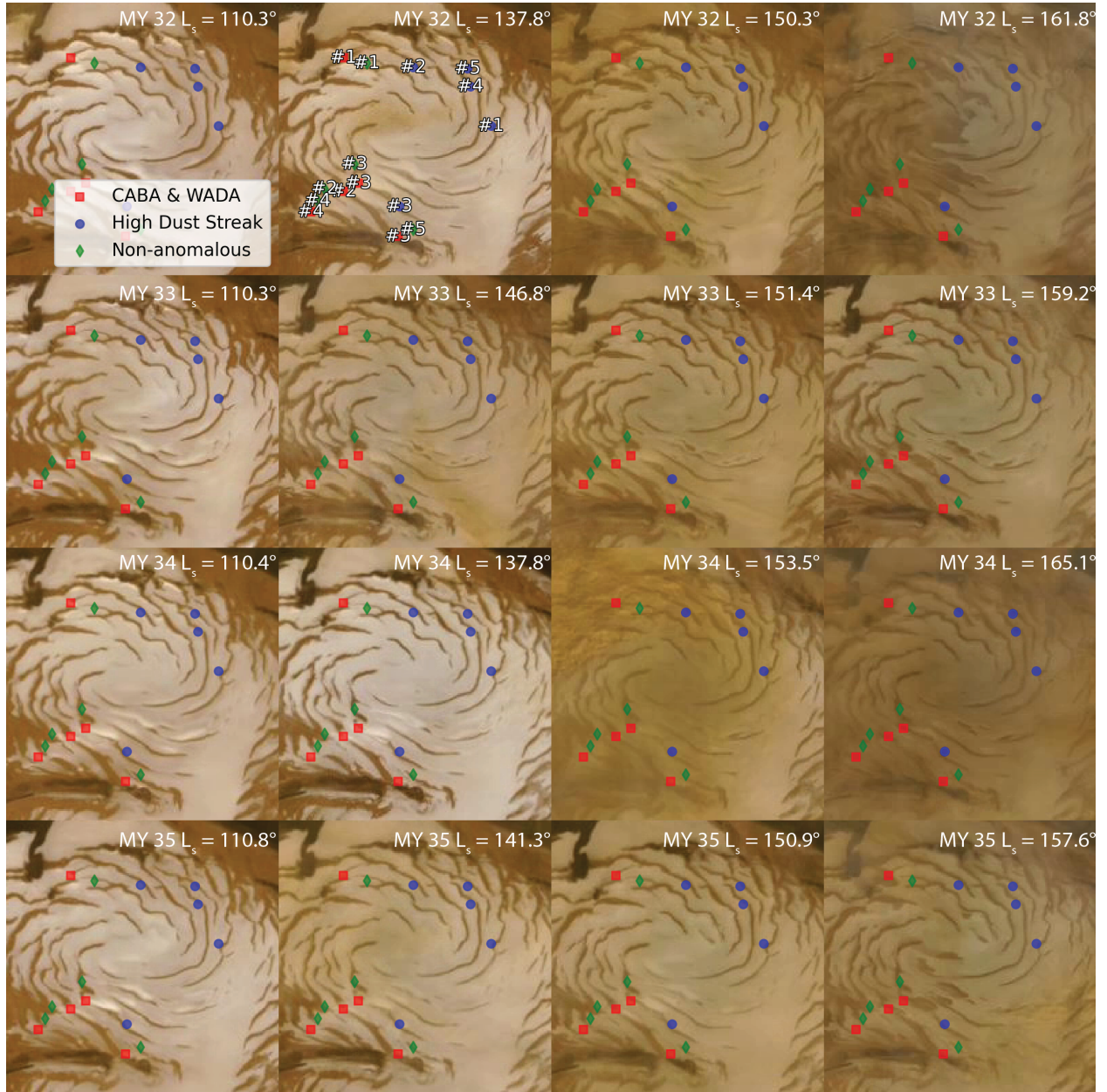


Figure 4.13: Similar to Figure 4.12 but showing the MARCI mosaic crops for MYs 32-35. The first column shows the CABA during early summer, the second column shows the first darkening event, the third column shows the second darkening event, and the fourth column shows the final darkening event. The WADA transformation follows similar trends, with the number and size of WADAs increasing with each darkening event. MY 32 has the largest darkening event, where about half of the NPRC undergoes the WADA transformation.

MARCI mosaic crop for MYs 32-35 sequences capturing key moments around the time of peak brightness difference (First Column), first darkening event (Second Column), second darkening (Third Column), and the third darkening event (Fourth Column).

4.3.2 Evolution Observed by THEMIS

Before $L_s = 82^\circ$, the temperature differences between the anomaly sites and their non-anomalous counterparts are negligible, mirroring the trends noted in the brightness (Figures 4.3, 4.7). Around $L_s = 82^\circ$, the temperature differences begin to diverge as the anomaly sites warm up more slowly than the non-anomalous counterparts. Similarly to the brightness divergence, the exact L_s date of temperature divergence varies, occurring as early as $L_s = 79^\circ$ and as late as $L_s = 83^\circ$. Limited data availability between $L_s = 70^\circ$ and $L_s = 82^\circ$ for the CABA and WADA site #1 complicates the determination of a precise divergence date. The temperature difference reaches an annual maximum at $L_s = 105^\circ$, where the anomaly sites are typically 15 K cooler than the non-anomaly sites. Notably, at CABA and WADA site #4, the temperature difference is significant at 30 K, whereas at CABA and WADA site #5, it is the smallest at 12 K. A parallel pattern emerges between the anomaly sites and high dust streak sites, with a temperature divergence date at $L_s = 82^\circ$ and an annual maximum difference of 33 K at $L_s = 105^\circ$. As CABAs transform into WADAs, most anomaly sites exhibit temperatures 3-4 K above their non-anomaly counterparts, consistent with earlier findings (Kieffer and Titus 2001). However, the limited availability of observations complicates the ability to precisely constrain this value.

Visual observations from THEMIS at the CABA sites show that anomaly sites are not only brighter but also display a smoother surface texture compared to their surroundings (Figure 4.4). As WADAs progress, a noticeable change in surface albedo occurs, distinguished by a prominent bright “halo” encircling the newly formed WADAs (Figure 4.4). A closer analysis of the halo reveals streaks oriented approximately perpendicular to the adjacent trough, abruptly ending at the onset of the trough. Additionally, the cloud map shows an absence of clouds at the CABA sites during spring, with the most significant cloud presence at the high dust streak sites (Figure 4.5).

4.3.3 Observations from HiRISE, CTX and MOLA

CTX observations, especially during periods of high-temperature differences, reveal notable smoothness at the CABA sites compared to their surroundings - seen in the THEMIS images as well (Figure 4.8). This smoothness manifests as a reduced presence of surface streaks at these anomaly sites. Concurrently, THEMIS observations of WADAs display a distinct bright halo outlining the newly formed WADAs, composed of small streaks that align roughly perpendicular to the adjacent trough (Figure 4.8). Some CTX observations also capture large-scale storms passing over the WADA sites (Figure 4.8). HiRISE observations of the CABA and WADA site #5 corroborate these findings, depicting the same halo structure identified in THEMIS and CTX imagery. HiRISE indicates that the halo streaks are roughly 650 meters in length and shows that during early summer, the surface at the CABA site is considerably smoother than at the WADA site which appears later.

The topographic profiles reveal that each anomaly site is located on a local topographic rise, typically 30 meters high (Figure 4.9). CABA and WADA site #1 is situated 150 meters above the surrounding terrain, while CABA and WADA site #5 is elevated by 10 meters. WADAs predominantly emerge around this local rise, with halos visible on either side. This distinctive local topography contrasts sharply with the non-anomalous sites, which do not have the local topographic rise found at the anomaly sites.

4.3.4 Surface Wind Evolution

To investigate wind behaviour, we assessed the relative differences in wind speeds between anomalous and non-anomalous sites as depicted in Figure 4.14. Before the peak katabatic wind season, up to $L_s = 70^\circ$, wind speeds at the anomaly sites are generally 10-25% weaker than those at non-anomaly sites. However, this pattern does not hold at CABA and WADA sites #1 and #3, where winds are 15-20% stronger than at corresponding non-anomaly locations. During the peak katabatic season, spanning $L_s = 70^\circ$ to $L_s = 90^\circ$, most anomaly sites continue to exhibit weaker

winds compared to non-anomaly sites, with differences ranging from 5% to 25%. Typically, the disparity in wind speeds diminishes as L_s dates progress within this interval. Contrarily, at CABA and WADA sites #3 and #5, the difference in wind speed increases with L_s dates. From $L_s = 90^\circ$ to $L_s = 120^\circ$, while most anomaly sites still experience significantly weaker winds than non-anomaly sites, on average 20% weaker; however, there are some sites such as CABA and WADA site #3 and CABA and WADA site #5 where the winds are about 10% faster than the respective non-anomaly site. After $L_s = 120^\circ$, all anomalous sites exhibit slower winds compared to non-anomalous sites, with the difference reaching up to 25%. During darkening events, wind speeds are simulated to exceed 20 meters per second at all sites (Figure 4.10). Notably, wind speeds at anomalous sites are comparable to those at non-anomalous sites but are up to 25% faster than those at sites with high dust streak activity. The variability in wind speed differences between anomaly and non-anomaly sites might be influenced by limitations in spatial resolution, particularly notable at CABA and WADA site #5, which is situated on a local elevation of only 10 meters—the smallest among all studied sites. The current resolution constraints of the measurement methods do not adequately capture the local topographic effects at such minor elevations, indicating a need for simulations with finer resolution.

A close examination of late summer eddies reveals that strong eddies often originate at the edge of the simulated domain (Figure 4.15). These eddies rapidly expand in both size and intensity within a single Martian day, covering hundreds of kilometers in about four hours. They approach the edge of the NPLD within the longitudinal range of 180° E to 330° E, aligning approximately with the general distribution of WADAs (Figure 4.15). Additionally, we note a significant decrease in the strength of these eddies between $L_s = 145^\circ$ and $L_s = 155^\circ$, coinciding with the period between the first and second darkening events.

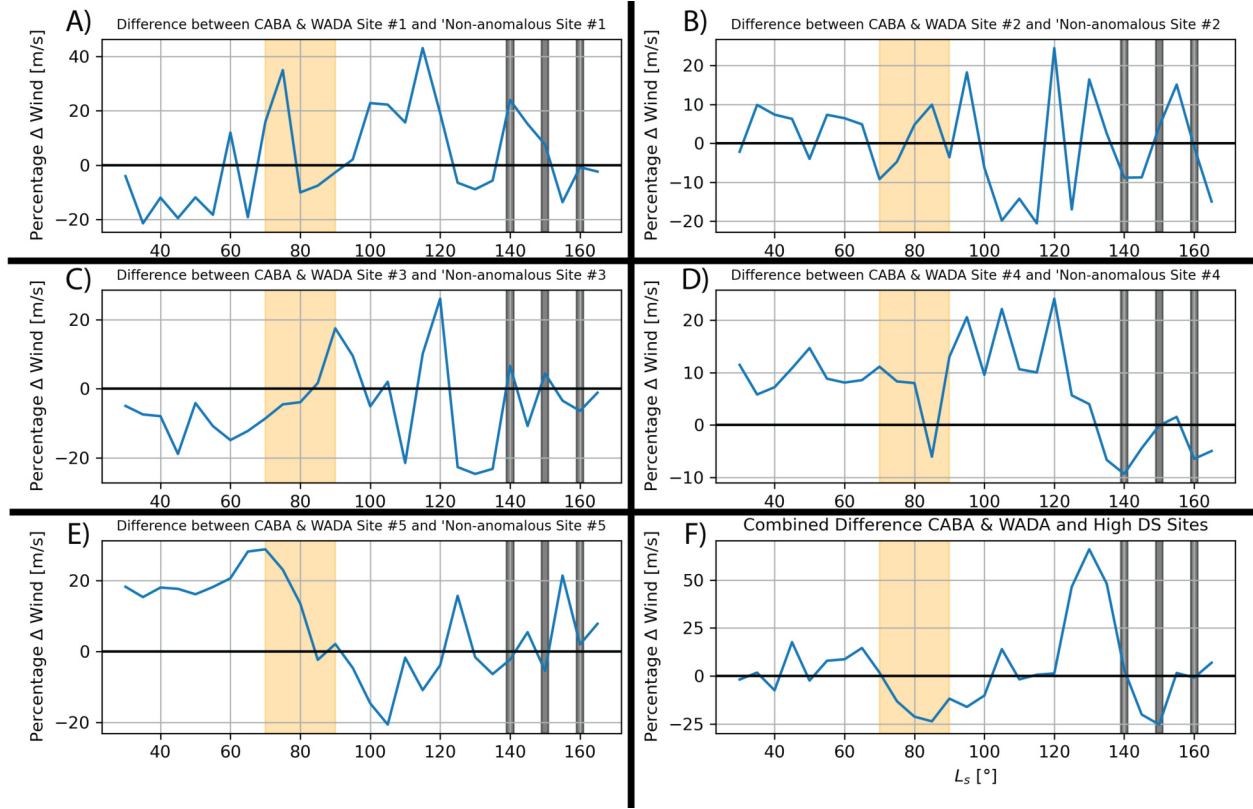


Figure 4.14: Percentage difference in wind speeds between anomalous sites and their corresponding non-anomalous and high dust streak sites. Positive values indicate slower wind speeds at the anomaly sites compared to the others. The x-axis represents the L_s [°] and the y-axis represents the difference (Δ) in max velocity in the simulated L_s date. The orange-shaded area represents the peak katabatic wind season and three dark lines mark the major darkening events. It is observed that most anomaly sites have weaker winds before and after the peak of the katabatic wind season. However, wind speeds at these anomaly sites significantly increase during late summer, coinciding with the occurrence of WADAs.

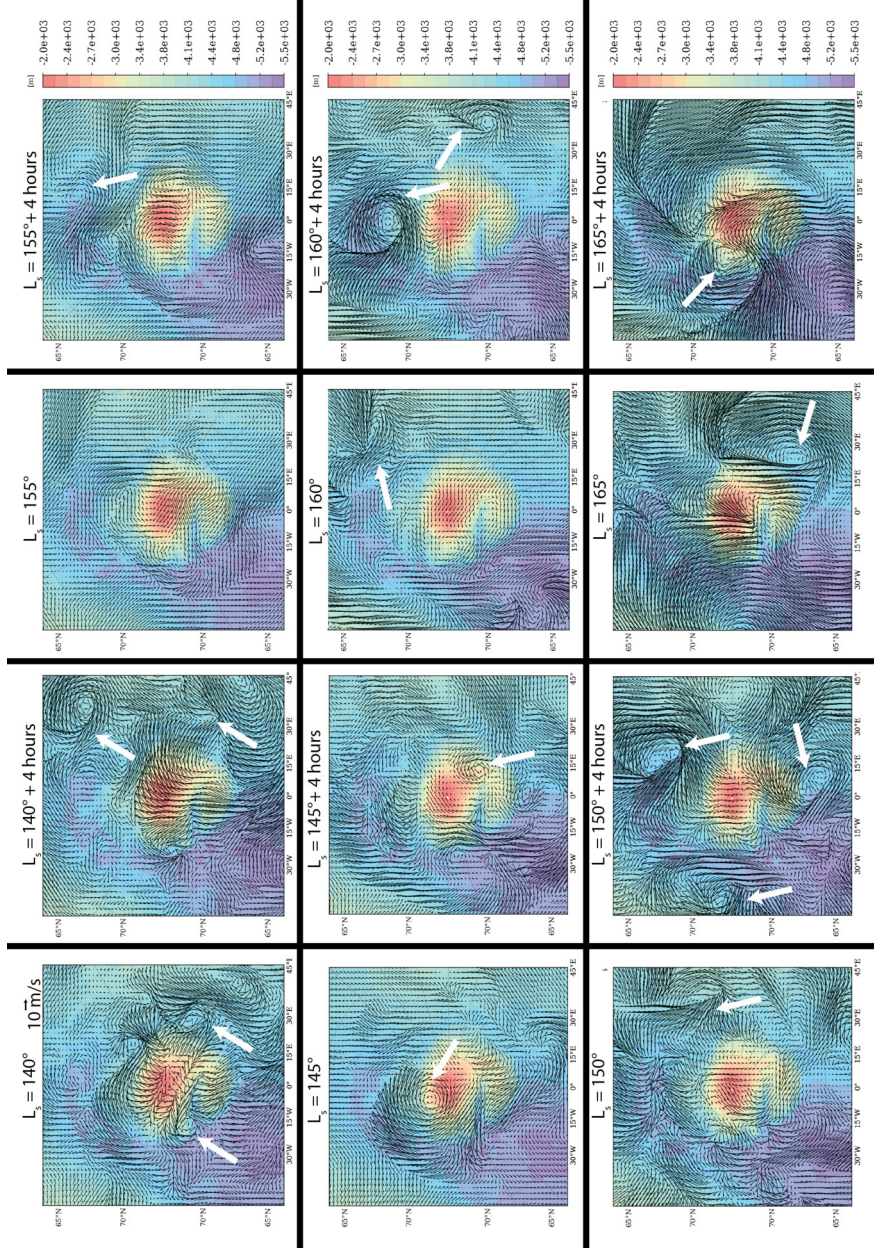


Figure 4.15: Late-summer MMM outputs from $L_s = 140^\circ$ to $L_s = 165^\circ$, showing wind vectors (black arrows) that are proportional to wind velocity and overlaid on the topography. White arrows highlight some of the larger eddies within the domain. These eddies traverse a significant portion of the domain within four hours, with peak activity during $L_s = 150^\circ$ to $L_s = 165^\circ$, particularly evident in the fifth and sixth rows of the figure. The timing of these observations corresponds to after the first and second darkening events at $L_s = 145^\circ$ and $L_s = 155^\circ$, respectively.

4.4 Discussion

4.4.1 Formation of the Cold and Bright Anomalies

Kieffer and Titus (2001) suggest that surface temperatures at the Vostok site increase more slowly due to cold trapping. This cold-trapping process creates a positive feedback loop, limiting solar absorption and maintaining a colder and brighter surface compared to the surroundings. However, cold trapping necessitates an initially cold surface to start the loop. Sites on the NPRC might retain their seasonal frost longer than their surroundings for several reasons, such as elevation effects and surface thermal inertia (Wood and Paige 1992; Blackburn et al. 2010). Calvin and Titus (2008) observed no major elevation cooling effect occurring at the Vostok anomaly site and the surrounding regions, leading to the conclusion that cold trapping is driving the albedo difference during early summer. Also, thermal inertia differences are expected to be minimal as the anomaly sites and the surrounding area show no discernible difference.

We propose that anomaly sites may form during late spring through aeolian processes, which involve the movement and interact of the lower atmosphere and the surface. Aeolian processes encompass erosion, transportation, and deposition of material by near-surface winds such as katabatic winds. We hypothesize that CABA sites are locations on the NPRC that experience weaker katabatic winds. Driven by topography, katabatic winds, which increase in speed as they flow downhill due to gravitational pull, typically enhance local sublimation rates. This enhancement occurs as these winds transfer significant kinetic energy and momentum to the surface, intensifying the energy exchange processes that drive the sublimation of ice directly to vapour (Smith et al. 2015; Smith and Spiga 2018). If CABA sites experience weaker katabatic winds, these sites would retain seasonal frost for a longer period, fulfilling the conditions needed to initiate cold trapping.

Several observations indicate these winds may be responsible for the appearance of CABAs and CABA sites experience weaker winds. The divergence in the temperature and brightness evolution during peak katabatic winds season suggests the role of these winds, as these winds are very

influential during this period (Smith et al. 2013; Smith and Spiga 2018). Additionally, cloud maps show that surface clouds associated with katabatic winds are less common at the anomaly sites compared to non-anomaly sites. Wind models from Smith and Spiga (2018) reveal that winds are generally weaker at the anomaly sites until $L_s = 80^\circ$. We suggest that the winds are weaker at these sites due to a local topographic rise found only at the anomaly sites (Figure 4.16). As katabatic winds flow over these sites, the winds slow down, converting kinetic energy into potential energy. The reduced wind speeds likely decrease local sublimation rates compared to non-anomalous sites that lack this characteristic topography, potentially initiating the cold trapping process. This effect of weaker katabatic winds persists into early summer allowing for the cold trapping process proposed by Kieffer and Titus (2001).

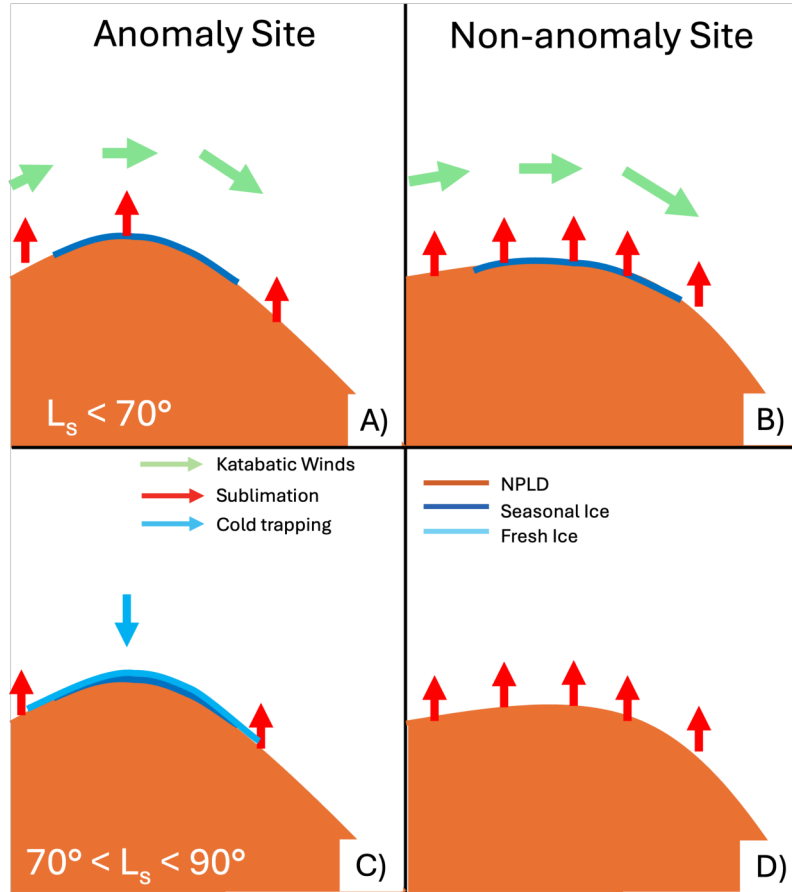


Figure 4.16: An illustrative sequence (not to scale) depicting the changes at an anomaly site (A, C) and a non-anomaly site (B, D) during late spring and early summer. The diagram's layers are colour-coded: orange for the dirty ice/sub-surface of the NPRC, dark blue for seasonal ice, and light blue for small-grain ice. Wind and atmospheric processes are represented by arrows: green for katabatic winds, red for sublimation, and light blue for cold trapping. Katabatic winds, slowed by the local topographic rise, lead to reduced sublimation rates at anomaly sites compared to non-anomaly sites. The preserved seasonal ice at anomaly sites acts as a cold trap for small-grain water ice, while non-anomaly sites lose most of their seasonal ice.

During early summer, the temperature and brightness difference continue to grow due to the cold trapping process. Notably, this process continues until $L_s = 108^\circ$, when both differences reach their respective annual peak. From $L_s = 80^\circ$ to $L_s = 108^\circ$, models show significantly slower winds at the anomaly sites, which we suggest may facilitate cold trapping; however, these connections need to be analyzed in more detail. We theorize that rapid winds may disrupt the deposition of fine-grained ice on the surface, thereby impeding the cold trapping process. From the annual peak until $L_s = 130^\circ$, the difference trends towards zero due to the general brightening of the NPRC where the

entire residual cap is covered by fine-grained water-ice (Calvin et al. 2015; Brown et al. 2016)

4.4.2 Formation of the Warm and Dark Anomalies

During late summer, after $L_s = 120^\circ$, late-summer transient eddies have been seen rapidly moving across the region. Tyler and Barnes (2005) have simulated these eddies to be hundreds of kilometers wide at an altitude of 25 kilometers, which can traverse the entire NPRC within a few hours - as seen in our late summer simulations. These eddies undergo substantial growth in both size and intensity with L_s . Hinson et al. (2022) have demonstrated that transient eddies, which are temporary and localized disturbances in the Martian atmosphere, play a critical role in the formation of the large spiral clouds observed annually. These eddies contribute to vertical atmospheric mixing and enhance the cooling effects of nocturnal water-ice aerosols, leading to convective instability. This process is crucial for the development of Nocturnal Mixed Layers that are closely associated with these cloud formations. Such dynamics consistently occur around the same $L_s = 120^\circ \pm 5^\circ$, each year, facilitating the repeated emergence of these spiral clouds at specific geographical locations influenced by the topographical features of the region and the seasonal atmospheric conditions. Cantor et al. (2010) noted three major darkening events at approximately at $L_s = 140^\circ$, 150° , and 160° - similar to MARCI observations from MY 29-35. This suggests that there is a strong correlation between the WADAs and the transient eddies, particularly in terms of the repeatability, size, and speed of the darkening events and the transient eddies. Additionally, spectral work from Cantor et al. (2010) showed that the Vostok site after the removal of high albedo material is dominated by water-ice.

Moreover, several observations indicate an aeolian mechanism behind the formation of WADAs. Firstly, we observe large dust storms that coincide with the appearance of WADAs, which we believe to be a part of these eddies. Secondly, the small streaks composing the halos indicate wind-shear-like interactions occurring at the surface. Thirdly, the position of the halos, present on either side of the characteristic topographic rise at the anomaly site, suggests hints at wind-shear-like interactions. Finally, our mesoscale models show a major decrease in the strength of these eddies

during $L_s = 145^\circ$ and $L_s = 155^\circ$, the period between the darkening event, further suggesting a link between the eddies and these anomalies.

We hypothesize as the eddies move NPRC, the local topographic rise at the anomaly sites results in an accelerated near-surface flow due to the topographic multiplier effect (Holmes et al. 2005) (Figure 4.17). This accelerated flow will enhance shear rates at these sites and remove the fresh, high albedo surface ice and expose water ice with dust inclusions, or "dirty" ice. It has been shown that dust inclusions less than 1% can reduce the albedo of water-ice from 1.0 to 0.1 (Khuller et al. 2021). The darker ice will have more solar absorption resulting in a warmer temperature we observe at the WADA sites. The halos are regions where the winds are insufficient at removing the surface ice, creating a stark albedo contrast between the brighter, purer ice and the dirty ice.

To explain the interannual variability in the number and size of WADAs we observe; this hypothesis suggests that the shear velocity in some MYs is higher than in others. For an average MY, only the fresh, bright CABA material is removed. During MYs with strong eddies, much larger WADA areas can be stripped of their summer frost.

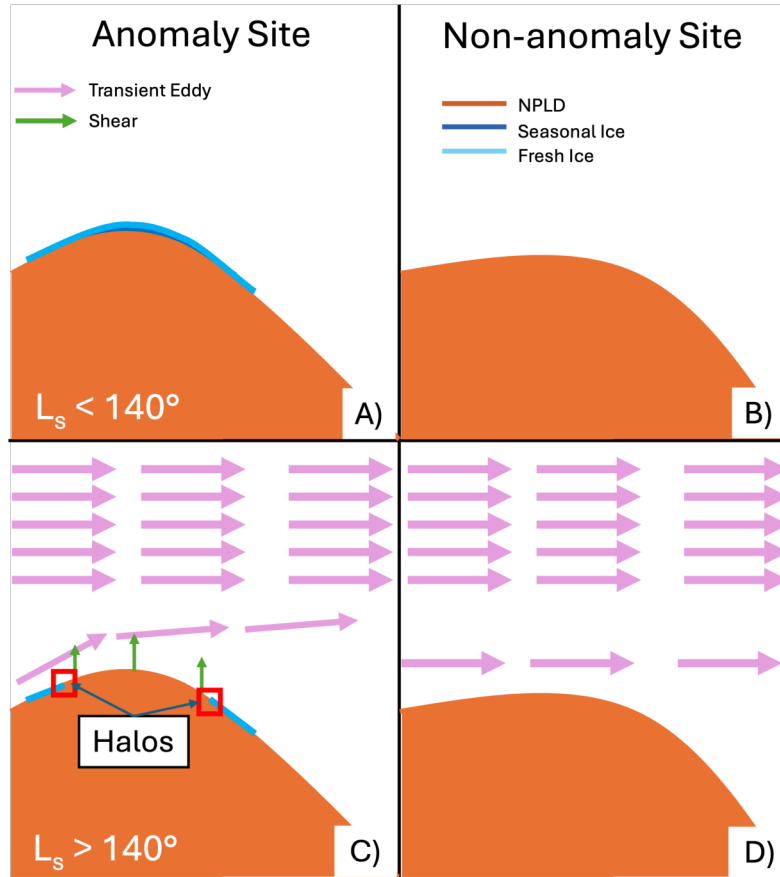


Figure 4.17: An illustrative sequence (not to scale) depicting the changes at an anomaly site (A, C) and a non-anomaly site (B, D) during summer and late summer. The diagram’s layers are colour-coded: orange for the dirty ice/sub-surface of the NPRC, dark blue for seasonal ice, and light blue for small-grain ice. Wind and atmospheric processes are represented by arrows: pink for late-summer transient eddies and purple for shear due to transient eddies. During summer, both site types exhibit similar albedo ice coverage. During late summer, transient eddies, influenced by the topographic effect multiplier, generate faster winds near the rise peaks at anomaly sites. This results in the removal of small-grain ice and potential downwind transport, leaving behind high albedo ice on both sides of the rise and exposing the underlying dirty ice.

4.5 Conclusion/Summary

In this chapter, we utilized data from multiple instruments—MARCI, THEMIS, CTX, HiRISE, and MOLA—to analyze five anomaly sites, five nearby non-anomaly sites, and five sites characterized by intense surface winds. This approach was aimed at understanding the processes potentially driving the formation of these anomalies. Additionally, the MMM was employed to examine the behaviour of seasonal reoccurring late-summer transient eddies.

MARCI surface albedo observations and THEMIS surface temperature measurements indicate that the CABA sites start to manifest around $L_s = 80^\circ$, aligning with peak katabatic wind speeds. At this stage, CABA sites become noticeably colder and brighter compared to both the non-anomaly and high dust streak sites, with the difference increasing until $L_s = 110^\circ$. This finding is consistent with prior research. Observations from THEMIS and CTX suggest that the surfaces at anomaly sites are smoother than their surroundings, while MOLA topography data reveal that all anomaly sites are located on local topographic rises ranging from 10 to 150 meters, features not present at the non-anomaly sites. Additionally, THEMIS cloud coverage data suggest that CABAs experience fewer elongated clouds, which typically form from katabatic winds, compared to the non-anomaly sites.

As late summer approaches, post- $L_s = 140^\circ$, MARCI detects three significant darkening events around $L_s = 140^\circ$, 150° , and 160° , marking the transformation of CABA sites into WADA sites. These transitions are accompanied by a 38% decrease in albedo and a temperature increase of approximately 5 K above that of non-anomaly sites, with each event intensifying the number and size of WADA sites. Notably, WADAs show more variability in terms of number and timing, occurring $\pm 5^\circ L_s$ dates from the previously mentioned L_s dates. These darkening events coincide with intense dust storms near the newly formed WADA sites and align with the late-summer transient eddies. Detailed observations from THEMIS, CTX, and HiRISE illustrate that WADAs are delineated by a bright halo on either side of the local rise, consisting of small streaks approximately 650 meters long, aligned perpendicular to nearby troughs.

We propose that CABAs are locations on the NPRC experiencing weaker katabatic winds relative to non-anomaly and high dust streak sites. These reduced winds lead to lower sublimation rates, maintaining the temperature differences observed in late spring. As the katabatic wind season concludes, these sites preserve the temperature difference due to cold trapping, where local cold spots condense highly reflective ice on the surface, forming CABAs as observed in thermal and visual data. The emergence of WADAs involves interactions between local topography and late-summer transient eddies. The local topographic multiplier effect, analogous to Bernoulli's Principle, accel-

erates near-surface winds as these large eddies encounter the topographic rises, increasing the shear rates and exposing an underlying layer of dirty ice. Halos form on either side of the rise, where the winds are insufficient to displace the surface ice. In instances where non-CABA sites transition into WADAs, it is believed that unusually strong late-summer transient eddies are the driving force.

Chapter 5

Summary, and Future Work

The northern and southern polar regions of Mars, extending approximately 40° from the respective geographic poles, are some of the most influential and dynamic regions of the planet. These regions are characterized by large seasonal CO_2 ice caps, residual water-ice caps, sub-surface layer deposits, and intense seasonal activities, which are crucial for understanding the present and ancient Martian climate. This dissertation has enhanced our understanding of the polar regions by introducing new sub-seasonal trends seen in the retreat of the seasonal caps, the impact of the GDS events, and providing the first hypothesis for CABAs and WADAs. This chapter of the dissertation summarizes the methods and major findings of previous chapters as well as proposes directions for future research.

5.1 Summary of Key Findings

5.1.1 Chapter 2: Tracking the North Polar Seasonal Cap Retreat of Mars Using Computer Vision

Using daily mosaics from the Mars Color Imager (MARCI) and a novel computer vision technique, we track the retreat of the North Polar Residual Cap (NPRC) from early northern spring to late spring over Mars Years (MY) 29-35. The computer vision method employs the HSV colour

format to accurately outline the NPSC and fit an ellipse. Compared with previous approaches, this automated method shows minor differences, less than 3° of latitude, primarily due to the higher degrees of freedom allowed by the ellipse fit as opposed to a circular fit. Additionally, the automated method is faster than the previous manual method, within two minutes using the automated method compared to weeks using manual methods. The automated method also utilized over 90% of the mosaics and produced a recession curve with higher temporal resolution.

The resulting recession curve indicates that the cap's retreat begins slowly and then accelerates towards the end of the season, a pattern we associate with enhanced sublimation driven by off-cap winds during this period. We also identify new sub-seasonal trends, such as a sudden increase in latitude during MY 30, which we attribute to a major sublimation event. Seasonal variations in the size and retreat rates of the NPSC are also observed following the MY 28 and MY 34 GDS events. Notably, the NPSC during MY 29 is smaller and sublimates slower than the seasonal average, whereas during MY 35, the NPSC is larger and sublimates faster. We propose that the timing of the GDS events critically influences the behaviour of the NPSC.

5.1.2 Chapter 3: Tracking the South Polar Seasonal Cap Retreat of Mars Using Computer Vision

Using daily mosaics from the MARCI and an enhanced computer vision technique, we track the SPSC from early southern spring to southern summer over MY 29-31. This method improves the static colour detector discussed in Chapter Two by integrating a dynamic algorithm that more effectively tracks the SPSC through regional dust storms. Compared with previous methods, the automated approach exhibits minor differences, likely due to variations in spatial resolution and the use of ellipse fits rather than circular fits. However, the automated method is considerably faster than earlier methods using the MARCI dataset and utilizes over 90% of the observations, yielding a recession curve with higher temporal resolution. We also monitor the recession of the MM seasonal cap outlier for the same MYs.

The recession curve demonstrates significant variations, with periods of acceleration and decel-

eration that correlate with changes in seasonal activities. Additionally, we examine the impact of the MY 28 GDS event on the SPSC and MM seasonal cap recession. We find after the MY 28 GDS event, the SPSC is significantly smaller and more advanced than the seasonal average, a trend also observed in the NPSC retreat during MY 29, a post-GDS year. For the MM seasonal cap, we note that the seasonal cap outlier is significantly smaller and sublimates faster following the GDS event. Interestingly, the MM seasonal cap is larger than average before the onset of the MY 28 GDS event, suggesting that pre-storm conditions may also influence cap dynamics.

5.1.3 Chapter 4: The formation and evolution of the Cool and Bright Anomalies and Warm and Dark Anomalies on the North Polar Residual Cap

Using thermal, visual, and topographic observations, we characterize the evolution and surface characteristics of five CABA and WADA anomaly sites (including four newly observed), five non-anomaly sites, and five high dust streak sites. Additionally, we employ a mesoscale model developed by LMD to simulate surface winds at these sites from early spring to late summer.

We observe that anomaly sites begin to diverge from non-anomaly sites between $L_s = 75^\circ$ and $L_s = 82^\circ$, coinciding with peak katabatic wind speed and associated atmospheric phenomena such as near-surface clouds. Interestingly, anomaly sites exhibit weaker katabatic winds, as evidenced by the absence of clouds at these sites and supported by the mesoscale model simulations. The differences peak at $L_s = 108^\circ$, where anomaly sites are notably brighter and 30 K colder than their non-anomaly counterparts. Topographic data reveals that anomaly sites are situated on local elevations ranging from 10 to 150 meters. During late summer, we document three major darkening events at these sites, where the albedo decreases by 38%, and temperatures are 5 K warmer than at non-anomaly sites, signalling the transition from CABA sites to WADA sites. These transitions are marked by large spiral dust storms that resemble eddies, and a distinct bright halo outlines the WADA sites situated on either side of the local elevations. This halo consists of small streaks aligned perpendicularly to the adjacent troughs. Furthermore, we find that the seasonal variability of these darkening events closely aligns with the activity of late-summer transient eddies.

We propose that CABAs form at sites on the NPRC that experience weaker katabatic winds, enabling these sites to retain seasonal frost longer than surrounding areas. During early summer, this difference is enhanced by cold trapping, allowing these sites to remain cooler. Conversely, WADAs appear to form as late-summer transient eddies rapidly traverse the surface, eroding the surface material and exposing underlying darker "dirty" ice. The halos are areas where surface wind speeds are insufficient to remove the surface ice.

5.2 Future Work

5.2.1 Future Research related to Chapter 2

An area of future work involves testing and verifying the proposed mechanism for the sudden latitude increase observed during MY 30. This event has been documented solely through observations from the Mars Color Imager (MARCI); however, employing spectral instruments during this event could offer deeper insights into the underlying processes. Spectral instruments, capable of distinguishing between CO₂ and water ice, may reveal the behaviour of these different ices during the event and their impacts on the sudden latitude increase. Additionally, using mesoscale models (resolution <10 km) to simulate surface winds during this event could enhance our understanding of the complex atmospheric dynamics near the edge of the NPSC and provide insights into surface wind speeds and temperatures.

Further research is also needed to understand the impact of the MY 28 and MY 34 GDS events by simulating these events and monitoring the NPSC before the MARCI observational period. Both GDS events occur during the northern winter, before the start of the MARCI dataset, making it impossible to observe the NPSC response with MARCI during these periods. However, studying the NPSC's behaviour during the GDS events could provide crucial insights into the processes influencing the formation and retreat of the NPSC. Spectral observations and weather simulations during these periods could simulate the movement of dust within the northern polar region, surface temperature, and other atmospheric processes, thereby confirming the mechanisms proposed in

Chapter 2.

Another prospective area of research builds on the findings from Chapter 2 by integrating the high-resolution outlines of the NPSC into weather models, particularly mesoscale models with a resolution finer than 1 kilometer. Currently, the mesoscale model developed by Le Laboratoire de Météorologie Dynamique assesses the NPSC extent using thermal observations with a resolution of 60 kilometers (Smith and Spiga 2018) or by calculating surface temperature; if the temperature is below the CO₂ condensation point, it is marked as NPSC. Incorporating these high-resolution outlines, with a pixel scale of 1.9 kilometers, could significantly enhance the model's ability to delineate the complex interactions occurring at the edge of the cap.

The final proposed area of future work is the continued monitoring of the NPSC retreat beyond MY 35. As MY 37 is concluding, the data necessary to create MARCI mosaics will soon be available. Continual monitoring of the NPSC retreat will provide an accurate depiction of the NPSC retreat, enabling the identification of anomalous MYs. Additionally, the high-resolution recession curves provided by the autonomous method will allow for a more detailed characterization of seasonal trends.

5.2.2 Future Research related to Chapter Three

The future research directions related to Chapter 3 align closely with those proposed for Chapter 2.

One area for future work is the continued monitoring of the SPSC retreat beyond MY 31. As MY 37 is drawing to a close, the data necessary to create MARCI mosaics will soon be available, leaving six seasons' worth of observations yet to be analyzed, including the MY 34 GDS event. Analyzing additional Mars Years will provide a more robust baseline average that can identify recurring sub-seasonal trends and anomalous MYs. Additionally, tracking the SPSC during MY 34 could provide insights into how the SPSC responds to GDS events and whether its behaviour parallels that of the NPSC.

Another similar area for future research involves the incorporation of high-resolution SPSC

outlines into mesoscale modelling. This endeavour, paralleling the motivation discussed previously, aims to utilize high-resolution maps combined with finer spatial resolution to enhance the model's ability to simulate complex interactions occurring at the edge of the cap.

Further research is also needed to understand the impact of the enlarged MM seasonal cap outlier prior to the onset of the MY 28 GDS event. This research would involve characterizing the MM seasonal cap for other GDS years such as MY 25, 28, and 34 to determine whether the results observed in this chapter are consistent across other GDS years. Additionally, this research will involve using a weather model to simulate the impact of the enlarged outlier on the storms prior to the onset of the GDS events, determining whether the outlier can serve as an early indicator for these events. Finally, the research aims to understand the potential conditions and processes required for an enlarged MM seasonal cap prior to the onset of these events. If the enlarged outlier is indeed a potential early indicator, it could provide insights into the forces that drive the mysterious GDS events and enhance our ability to predict these storms.

5.2.3 Future Research related to Chapter 4

The future research directions related to Chapter 4 are centred on validating the proposed hypotheses and addressing the remaining questions.

Currently, the mesoscale models utilized are unable to fully resolve the interactions between the local topographic rise at the anomaly sites and the near-surface winds. To confirm the interactions proposed in the chapter, fine-resolution simulations with a spatial resolution finer than 3 kilometers are necessary.

Another area for future research involves the continual monitoring of the anomaly sites and expanding the number of observed sites. Chapter 4 has identified four new anomaly sites, beyond Vostok, yet numerous other locations on the NPRC exhibit characteristics of CABAs and WADAs. Due to limited observations, only five anomaly sites could be characterized. Studying more sites will enhance our understanding of the differences between anomaly and non-anomaly sites and will refine the existing hypotheses. Moreover, continual monitoring will provide additional observations

during the late-summer transition, as current data are sparse concerning the transition of CABA sites into WADA sites.

The final area for future work is the spectral analysis of the anomaly sites to determine the grain size at the CABA sites and the spectral composition at the WADA sites. It has been proposed that CABAs are brighter due to cold trapping, characterized by small-grained water-ice. Spectral modelling and analysis may provide constraints on the grain size necessary to achieve the observed albedo, thus offering further insights into the CABA formation process. We hypothesize that the WADA sites feature exposed dirty ice. Spectral modelling and analysis will help determine the exact composition of the WADA sites, the amount of dust inclusion required to achieve the measured albedo, and more insight into the exact processes behind these anomalies.

Chapter 6

Appendix

6.1 Chapter 2: Appedenx

6.1.1 Polynomial Fits for the NPSC

Table 6.1: Table containing the coefficients for the linear fit and the 7th order fit for the NPSC latitude retreat

MYs	L _s start date [°]	1st order fit coefficients m [° per L _s], b [°]	7th order fit coefficients ($a_7, a_6, \dots, a_1, a_0$)
7-year average	0.5	0.23, 57.88	$6.94 \times 10^{-11}, -1.56 \times 10^{-8}, 1.32 \times 10^{-6},$ $-5.24 \times 10^{-5}, 8.27 \times 10^{-4}, 1.44 \times 10^{-3},$ $5.84 \times 10^{-2}, 6.00 \times 10^1$
29	24.9	0.23, 58.50	$9.70 \times 10^{-10}, -3.10 \times 10^{-7}, 4.13 \times 10^{-5},$ $-2.99 \times 10^{-3}, 1.27 \times 10^{-1}, -3.14,$ $4.23 \times 10^1, -1.76 \times 10^2$
30	24.5	0.24, 57.38	$1.40 \times 10^{-9}, -5.10 \times 10^{-7}, 7.81 \times 10^{-5},$ $-6.53 \times 10^{-3}, 3.20 \times 10^{-1}, -9.17,$ $1.42 \times 10^2, -8.55 \times 10^2$
31	20.7	0.25, 57.40	$-5.19 \times 10^{-10}, 1.65 \times 10^{-7}, -2.17 \times 10^{-5},$ $1.53 \times 10^{-3}, -6.29 \times 10^{-2}, 1.49,$ $-1.88 \times 10^1, 1.60 \times 10^2$
32	12.2	0.23, 58.02	$1.64 \times 10^{-10}, -4.57 \times 10^{-8}, -5.30 \times 10^{-6},$ $-3.32 \times 10^{-4}, 1.21 \times 10^{-2}, -2.56 \times 10^{-1},$ $3.10, 4.61 \times 10^1$
33	17.9	0.24, 57.41	$-1.02 \times 10^{-10}, 3.23 \times 10^{-8}, -4.19 \times 10^{-6},$ $2.88 \times 10^{-4}, -1.12 \times 10^{-2}, 2.47 \times 10^{-1},$ $-2.62, 7.18 \times 10^1$
34	0.5	0.23, 58.24	$6.48 \times 10^{-11}, -1.49 \times 10^{-8}, 1.37 \times 10^{-6},$ $-6.15 \times 10^{-5}, 1.41 \times 10^{-3}, -1.41 \times 10^{-2},$ $2.19 \times 10^{-1}, 5.97 \times 10^1$
35	6.9	0.23, 57.77	$7.43 \times 10^{-11}, -1.69 \times 10^{-8}, 1.46 \times 10^{-6},$ $-6.02 \times 10^{-5}, 1.16 \times 10^{-3}, -8.60 \times 10^{-3},$ $2.19 \times 10^1, 5.86 \times 10^1$

Table 6.2: Table containing the coefficients for the linear fit and the 7th order fit for the NPSC area retreat.

MYs	L _s start date [°]	1st order fit coefficients m [km ² per L _s], b [km ²]	7th order fit coefficients ($a_7, a_6, \dots, a_1, a_0$)
7-year average	0.5	$-1.12 \times 10^5, 1.02 \times 10^7$	$-3.06 \times 10^{-5}, 6.83 \times 10^{-3}, -5.66 \times 10^{-1},$ $2.03 \times 10^1, 2.16 \times 10^2, -3.89 \times 10^3,$ $-3.65 \times 10^4, 1.01 \times 10^7$
29	24.9	$-9.89 \times 10^4, 9.24 \times 10^5$	$-1.58 \times 10^{-4}, 4.80 \times 10^{-2}, -6.06,$ $4.10 \times 10^2, -1.60 \times 10^4, 3.16 \times 10^5,$ $-4.43 \times 10^6, 3.06 \times 10^7$
30	24.5	$-1.07 \times 10^5, 9.86 \times 10^6$	$-8.53 \times 10^{-4}, 3.07 \times 10^{-1}, -4.64 \times 10^1,$ $3.83 \times 10^3, -1.85 \times 10^5, 5.26 \times 10^6,$ $-8.08 \times 10^7, 5.26 \times 10^7$
31	20.7	$-1.11 \times 10^5, 1.00 \times 10^7$	$3.00 \times 10^{-4}, -0.56 \times 10^{-2}, 1.27 \times 10^1,$ $-9.05 \times 10^2, 3.75 \times 10^4, -9.02 \times 10^5,$ $1.15 \times 10^7, -5.22 \times 10^7$
32	12.2	$-1.01 \times 10^5, 1.01 \times 10^7$	$-1.05 \times 10^{-4}, 3.00 \times 10^{-2}, -3.56,$ $2.26 \times 10^2, -8.34 \times 10^3, 1.78 \times 10^5,$ $-2.13 \times 10^6, 1.95 \times 10^7$
33	17.9	$-1.12 \times 10^5, 1.02 \times 10^7$	$3.02 \times 10^{-5}, -1.00 \times 10^{-2}, 1.38,$ $-9.99 \times 10^1, 4.07 \times 10^3, -9.23 \times 10^4,$ $9.41 \times 10^5, 5.75 \times 10^6$
34	0.5	$-1.13 \times 10^6, 1.02 \times 10^7$	$-3.31 \times 10^{-5}, 7.90 \times 10^{-3}, -7.36 \times 10^{-1},$ $3.37 \times 10^1, -7.75 \times 10^2, 7.98 \times 10^3,$ $-1.42 \times 10^6, 1.03 \times 10^7$
35	6.9	$-1.14 \times 10^5, 1.05 \times 10^7$	$-4.35 \times 10^{-5}, 1.04 \times 10^{-2}, -9.84 \times 10^{-1},$ $4.66 \times 10^1, -1.19 \times 10^3, 1.77 \times 10^4,$ $-2.93 \times 10^5, 1.15 \times 10^7$

6.2 List of Conferences and Publications

6.2.1 Publications

- **Acharya, P. J.**, Smith, I. B., & Calvin, W. M. (2024). “Tracking the South Polar Seasonal Cap Retreat of Mars Using Computer Vision.” *Icarus*.
- **Acharya, P. J.**, Smith, I. B., & Calvin, W. M. (2023). “Tracking the Northern Seasonal Cap Retreat of Mars Using Computer Vision.” *Icarus*, vol. 390, Jan. 2023, p. 115295.
- Chojnacki, M., Vaz, D. A., **Acharya, P. J.**, Silvestro, S., & Smith, I. B. (2023). “Revelations of Interannual Dune Evolution from the Swiftest Aeolian System on Mars by MRO/HiRISE Long-Term Monitoring.” *Icarus*, pages 115863.
- Landis, M. E., **Acharya, P. J.**, Alsaeed, N. R., et al. (2023). “Polar Science Results from Mars Reconnaissance Orbiter: Multiwavelength, Multiyear Insights.” *Icarus*, pages 115794.
- Smith, I. B., Guzewich, S. D., Smith, D. E., Diniega, S., et al. including **Acharya, P. J.** (2020). “Solar-System-Wide Significance of Mars Polar Science.”

6.2.2 Conferences attended

- **Acharya, P. J.**, Smith, I. (2024). “Formation and Evolution of Cold and Bright Anomalies and Warm and Dark Anomalies on the North Polar Residual Cap.” Presented at *Eighth International Conference on Mars Polar Science and Exploration*, Jul. 2024
- **Acharya, P. J.**, Smith, I.B., & Calvin, W. (2024). “Tracking the Southern Polar Seasonal Cap Retreat of Mars Using Computer Vision.” Presented at *Tenth International Conference on Mars*, Jul. 2024
- **Acharya, P. J.**, Smith, I. B., & Calvin, W. (2022). “Interannual Variations in the Retreat of the Northern Seasonal Cap of Mars Using Computer Vision.” Presented at *Seventh International*

Workshop on the Mars Atmosphere: Modelling and Observations, Jun. 2022, 3306.

- **Acharya, P. J.**, & Smith, I. B. (2022). “Seasonal Variation of the Cold and Bright Anomalies on the North Polar Layered Deposits.” In *Seventh International Workshop on the Mars Atmosphere: Modelling and Observations*, Jun. 2022, page 3534.
- **Acharya, P. J.**, Smith, I.B., & Calvin, W. (2021). “Using Computer Vision to Monitor the Recession of the Northern Seasonal Cap.” Presented at the *AGU Fall Meeting Abstracts*, Dec. 2021, P35A-01.
- **Acharya, P. J.** & Smith, I. B. (2021). “Seasonal Variations of the Cold and Bright Anomalies on the North Polar Layered Deposits.” Presented at the *52nd Lunar and Planetary Science Conference*, Mar. 2021, 2541.
- **Acharya, P. J.** & Smith, I. B. (2021). “Seasonal Variation of the Cold and Bright Anomalies on the North Polar Residual CAP.” Presented at *AGU Fall Meeting* Dec 2020

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