

**Records of Ancient Climate Change on Mars:
Investigating Sedimentary Successions
in Gale Crater & Oxia Planum**

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Declaration of Originality

I certify that all work presented in this thesis, other than where clearly stated and referenced, is entirely my own and was conducted throughout the duration of this PhD at Imperial College London.

Amelie Roberts
September 2025

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“He was determined to discover the underlying logic behind the universe.

Which was going to be hard, because there wasn’t one.”

Terry Pratchett

Abstract

A central goal of Mars exploration has been to understand how the climate evolved from one that once supported liquid water to the cold, hyperarid desert observed today. This thesis examines the sedimentary geology and geomorphology of two regions on Mars from during and after the Noachian-Hesperian boundary, when surface water activity was thought to have waned. The overarching objective is to evaluate how the sedimentary environments of these regions evolved through time and the implications for regional climate change. The first study area, the basal section of the orbitally defined Layered Sulfate unit (LSu) in Gale crater, was explored in situ by the Curiosity rover. The second study area, the sediment fan at Oxia Planum, lies within one of the proposed landing ellipses of the upcoming Rosalind Franklin rover.

At Gale crater, rover and orbital datasets are used to investigate the lithofacies and sedimentary architecture of the basal section of the LSu. Two studies are presented. The first documents decametre-wide scour-and-fill structures, interpreted as aeolian, providing evidence for a period of intensified wind activity. The second examines the ~100 m basal succession, where the lithofacies are consistent with a predominantly aeolian environment, with thick sandsheets likely controlled by near-surface water. Together, these results suggest that groundwater and habitable conditions may have persisted longer than simple aridification models imply.

At Oxia Planum, high-resolution orbital datasets are used to reconstruct the geomorphology and photogeology of the sediment fan. The results reveal Re. These observations constrain the timing and longevity of aqueous activity in the Oxia basin and provide context for future rover exploration.

Collectively, this thesis demonstrates how late-stage aqueous processes shaped Martian sedimentary environments and highlights the importance of integrating rover and orbital datasets for understanding past habitability.

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ALR

Acronyms, Abbreviations, and Nomenclature

Acronyms and Abbreviations

APXS – Alpha **P**roton **X**-ray Spectrometer
BVS – Bela **V**ista **S**cour from Chapter 2
CaSSIS – Colour and Stereo Surface Imaging System
CDS – Canta **D**etour **S**cour from Chapter 2
CTX – Context Camera
ChemCam – **C**hemistry and **C**amera Complex.
CS1 – Chenapau **S**cour **1** from Chapter 2
CS2 – Chenapau **S**cour **2** from Chapter 2
CS3 – Chenapau **S**cour **3** from Chapter 2
DRT – Dust **R**emoval **T**ool (*Curiosity*)
ES1 – Exhumed **S**cour **1** from Chapter 2
ES2 – Exhumed **S**cour **2** from Chapter 2
ES3 – Exhumed **S**cour **3** from Chapter 2
ESA – European Space Agency
HiRISE – High **R**esolution **I**maging **S**cience **E**xperiment
IAU – International Astronomical Union
LSu – Layered **S**ulfate-bearing **u**nit
Mastcam – **M**ast **c**amera (*Curiosity*)
MAHLI – Mars **H**and **L**ens **I**mager (*Curiosity*)
MER – Mars **E**xploration **R**overs (*Opportunity and Spirit*)
MOLA – Mars **O**rbiter **L**aser **A**ltimeter
MSL – Mars **S**cience **L**aboratory
NASA – National Aeronautics and Space Administration
Navcam – Navigation **c**amera
OS1 – Orinoco **S**cour **1** from Chapter 2
OS2 – Orinoco **S**cour **2** from Chapter 2
THEMIS – Thermal **E**mission **I**maging **S**ystem

IAU Recognised Names

The International Astronomical Union (IAU) is responsible for assigning official nomenclature to features on Mars. Consequently, all other locations, landmarks, landforms and geologic or stratigraphic units that were informally named by myself, the MSL Team (Vasavada et al., 2014), or the ESA Rosalind Franklin rover team (Vago et al., 2017) are not capitalised.

Aeolis Mons

Aeolis Palus

Amazonis

Arabia Terra

Argyre Basin

Becquerel Crater

Chryse Planitia

Earth

Eberswalde

Elysium

Gale

Gediz Vallis

Hellas

Hesperis Terra

Isidis Basin

Juventae Chasma

Jezero

Kilkhampton

Ladon Vallis

Mars

Melas Chasma

Meridiani Planum

Noachis Terra

Oxia Planum

Terra Tyrrhena

Tharsis

Tharsis Rise

Utopia Basin

Valles Marineris

Table of Contents

Declaration of Originality	2
Copyright Declaration	3
Abstract.....	6
Acknowledgements	7
Acronyms, Abbreviations, and Nomenclature	9
List of Figures.....	16
List of Tables	19
Chapter 1: Introduction and Literature Review.....	20
1.1 Introduction.....	20
1.2 A Brief History of Evidence of Surface Water on Mars	20
1.2.1 Orbiters	20
1.2.2 Landers and Rovers.....	24
1.3 The Martian Climate Conundrum	26
1.4 Geologic history of Mars	28
1.4.1 Distinct Physiographic Controls on Mars	28
1.4.2 Introduction to Martian Geologic Time	29
1.4.3 Pre-Noachian Mars	30
1.4.4 Noachian Mars	30
1.4.5 Hesperian Mars	33
1.4.6 Amazonian Mars	34
1.5 Open Questions Regarding Martian Climate Change.....	35
1.6 Research Questions and Objectives	36
1.7 Statement of Originality and Contribution.....	37
1.8 Publications	37
Chapter 2: Palaeo-scours within the layered sulfate-bearing unit at Gale crater, Mars: Evidence for intense wind erosion.....	40
2.1 Introduction.....	40
2.1 Geologic and Stratigraphic Context.....	41
2.3 Data and Methods	45
2.3.1 Instruments.....	45
2.3.2 Data Collection and Imaging Strategy	46
2.3.3 Data analysis	46
2.4 Stratigraphic context of scour-and-fill structures.....	47
2.4.1 Observations	50
2.4.2 Interpretation.....	52
2.5 Description of Scours-and-fill Structures	53

2.5.1 Class A Scour-and-fill structures with a simple, symmetric morphology and light-toned infill	55
2.5.2 Class A Scour-and-fill structures with a complex morphology and light-toned infill	62
2.5.3 Class A scour-and-fill structures that have been exhumed and contain minimal to no preserved infill	67
2.5.4 Class B Scour-and-fill structures	72
2.6. Hypotheses for formation of scour-and-fill structures	74
2.6.1. Hypothesis 1: A fluvial mechanism for scour-and-fill structures	76
2.6.2. Hypothesis 2: Aeolian Bedform Migration as a Mechanism for Scour-and-fill Structures	77
2.6.3. Hypothesis 3: Scour formation by aeolian blowouts	78
2.6.4 Evolutionary model for scour-and-fill structure formation.....	81
2.7 Scour-and-fill structures as evidence for intensification of wind regime	83
2.8 Conclusions.....	84
Chapter 3: An Aeolian Depositional Sequence Shaped by Near-Surface Water at the Base of the Layered Sulfate Unit, Gale Crater, Mars	86
3.1 Introduction.....	86
3.2 Geologic Context	87
3.3 Methodology and data sets.....	88
3.3.1 Instruments and data products.....	88
3.3.2 Data Collection and Imaging Strategy	90
3.3.3 Data analysis	92
3.4 Stratigraphy and Sedimentology	93
3.4.1 Contigo Member	98
3.4.2 Catrimani member.....	103
3.4.3 Amapari Member	110
3.4.4 Chenapau Member	112
3.5 Discussion	133
3.5.1 What was the paleoenvironmental setting?	133
3.5.2 Implications.....	142
3.6 Conclusions.....	143
Chapter 4: Reconstructing Depositional Environments of the Sediment Fan in Oxia Planum, Mars	145
4.1 Introduction.....	145
4.2 Context.....	146
4.3 Methodology	148
4.3.1 Datasets	148
4.3.2 Mapping and Identification of Units	148
4.3.3 Unit Descriptions and Interpretations	149
4.4 Photogeologic Units.....	153
4.4.1 Basal units.....	153

4.4.2 Sediment Fan and Associated Units.....	160
4.4.3 Other Units.....	170
4.5 Morphology of the Sediment Fan	171
4.5.1 Channel-Basin Transition	171
4.5.2 Ribbon and Lobe-shaped Sedimentary Bodies	172
4.5.3 The distribution of proximal apices and their distance to the channel-basin transition	176
4.2.4 Sediment Fan Slope	177
4.5.5 The North-South Ridge.....	179
4.6 Discussion.....	181
4.6.1 History and timing of Emplacement	181
4.6.2 Towards a Working Hypothesis on the Fan Emplacement Model.....	182
4.6 Conclusions.....	188
Chapter 5: Synthesis, conclusions, and future work.....	190
5.1 Introduction.....	190
5.2 Records of ancient climate change on mars.....	190
5.2.1 The Layered Sulfate unit at Gale crater	190
5.2.2 The Sediment Fan at Oxia Planum.....	197
5.2.2 What do Oxia Planum and Gale crater tell us about possible climates during the mid-Noachian to early-Hesperian.....	204
5.3 Challenges in Understanding Climate Change on Mars: Image Resolution and Gaps in the Rock Record.....	204
5.4 Conclusions.....	205
5.5 Future Work.....	207
5.5.1 Investigating Spatial and Temporal Drivers of Dry Accumulation, Damp and Cementing Conditions, and Deflation in Aeolian Sandsheets.....	207
5.5.2 Global Survey on Late-stage Fan Reactivations on Mars.....	210
5.5.3 Testing Sedimentary Hypotheses with Upcoming Planetary Missions	211
Bibliography.....	215
Appendices.....	251
Appendix I. Figures	251
I.1 Map of key image locations used for stratigraphic mapping the LSu.	251
I.2 Map of stratigraphic contacts compared to the scour-and-fill structures.....	252
I.3. Example for manual measurements of apparent laminae thickness measurements.....	253
I.4. Measurements of apparent laminae thickness using peak detection on greyscale pixel intensities.	253
Appendix II. Tables	257
II.1 Complete table of missions to Mars	257
II.2 Table of Mastcam mosaics used in this work.....	259
II.3 Table of LD RMI mosaics used in this work.	260
II.4 Table of MAHLI mosaics used in this work.	260

II.5 Table of Scour Measurements Additional Information	261
II.6. Table showing the summary of stratigraphic unit elevations across major buttes and along the rover traverse within Marker Band valley.	261
Appendix III. Software and Techniques	263
III.1 Construction of DTMs and Orthomosaics	263
III.2 Construction of Viewsheds	263
III.3 White Colour Balancing.....	263
III.4 Measuring Dip and Strike	263
III.5 Data Availability	264

List of Figures

1.1	Selected evidence for water on Mars in Chapter 1.....	22
1.2	Triple point of water phase diagram and surface conditions on Mars & Earth in Chapter 1.....	27
1.3	Martian geologic timescales and terrestrial chronostratigraphy chart in Chapter 1.....	30
1.4	Global maps of Mars in Chapter 1.....	32
1.5	Schematic showing the geologic evolution of Mars in Chapter 1.....	35
2.1	Maps showing the study area at Gale crater in Chapter 2.....	43
2.2	Stratigraphic column of bedrock encountered by the Curiosity rover in Chapter 2.....	44
2.3	Maps showing the context of the scour-and-fill structures in Chapter 2.....	48
2.4	Mosaic showing the stratigraphic context of the scour-and-fill structures in Chapter 2.....	49
2.5	Examples of selected facies characteristic of the scour-enclosing strata in Chapter 2.....	51
2.6	Chenapau Scour 1 in Chapter 2.....	56
2.7	Additional details from Chenapau Scour 1 in Chapter 2.....	57
2.8	Chenapau Scour 2 in Chapter 2.....	59
2.9	Orinoco Scour 2 in Chapter 2.....	60
2.10	Chenapau Scour 3 in Chapter 2.....	61
2.11	Orinoco Scour 1 in Chapter 2.....	64
2.12	Additional details from Orinoco Scour 1 in Chapter 2.....	65
2.13	Bela Vista Scour in Chapter 2.....	66
2.14	Exhumed Scour 1 in Chapter 2.....	69
2.15	Exhumed Scour 2 in Chapter 2.....	70
2.16	Exhumed Scour 3 in Chapter 2.....	71
2.17	Canta Detour Scour in Chapter 2.....	74
2.18	Reconstructed architecture of each scour-and-fill structure to scale in Chapter 2.....	76
2.19	Schematic showing scenarios for scour-and-fill formation in Chapter 2.....	82
3.1	Geologic context used in planning in Chapter 3.....	91
3.2	The stratigraphic column of the Geologic formations, member, and units in Chapter 3.....	93
3.3	Selected examples of the stratigraphic architecture of unit A in Chapter 3.....	99
3.4	Selected examples of the facies of unit A in Chapter 3.....	100
3.5	Inferred boundary between the LSu and the underlying Clay-Sulfate transition in Chapter 3.....	101
3.6	Selected examples of the stratigraphic architecture of unit B in Chapter 3.....	104
3.7	Selected examples of the facies of unit B in Chapter 3.....	105
3.8	Selected outcrop expressions and lithofacies of unit C in Chapter 3.....	108
3.9	Selected outcrop expressions and lithofacies of unit D in Chapter 3.....	111
3.10	Selected outcrop expressions and lithofacies of unit E in Chapter 3.....	114

3.11 U-shaped scour embedded in unit E in Chapter 3.....	117
3.12 Selected outcrop expressions and lithofacies of unit F in Chapter 3.....	118
3.13 Selected outcrop expressions and lithofacies of unit G in Chapter 3.....	121
3.14 Selected outcrop expressions and lithofacies of unit H in Chapter 3.....	124
3.15 Examples of select facies characteristic of unit H in Chapter 3.....	125
3.16 Selected outcrop expressions and lithofacies of unit I in Chapter 3.....	128
3.17 Selected outcrop expressions and lithofacies of unit J in Chapter 3.....	131
3.18 Examples of select facies characteristic of unit J in Chapter 3.....	132
3.19 Example of unit K lithofacies in Chapter 3.....	133
3.20 Proposed spatial and temporal models for dry and damp sandsheet dynamics in Chapter 3.....	139
3.21 Geologic history of the basal section of the LSu in Chapter 3.....	141
4.1 Geographic context of the Oxia Planum sediment fan in Chapter 4.....	147
4.2 Photogeologic map of inferred stratigraphic units in Raetia Palus, Oxia Planum in Chapter 4.....	150
4.3 The Raised Linear Feature unit in Chapter 4.....	154
4.4 The Orthogonally Fractured unit in Chapter 4.....	156
4.5 The Pitted unit in Chapter 4.....	158
4.6 The Sub-polygonal Fractured unit in Chapter 4.....	159
4.7 The Upper Fractured unit in Chapter 4.....	161
4.8 The Ledge Forming unit in Chapter 4.....	163
4.9 The Medium-light Toned Smooth unit in Chapter 4.....	165
4.10 Internal stratification of the Medium-light Toned Smooth unit in Chapter 4.....	166
4.11 The Medium-light Toned Rough unit in Chapter 4.....	168
4.12 The Medium-light Toned Rough unit in Chapter 4 continued.....	169
4.13 Raised linear ridges on the surface of the Medium-light Toned Rough unit in Chapter 4.....	170
4.14 The Light-toned Crater Infill unit in Chapter 4.....	171
4.15 The channel-basin transition of the Oxia Planum sediment fan in Chapter 4.....	172
4.16 Craters embedded between layers of the sediment fan in Chapter 4.....	175
4.17 Correlation of sediment fan bodies in Chapter 4.....	176
4.18 Vertical displacement of the sediment fan by the north-south ridge in Chapter 4.....	178
4.19 The north-south ridges cross-cutting the sediment fan in Chapter 4.....	180
4.20 The fan emplacement model proposed for Oxia Planum in Chapter 4.....	187
5.1 Select facies of aeolian sandsheets in Meridiani Planum and Gale crater in Chapter 5.....	194
5.2 LSu in Gale crater compared to selected other layered deposits on Mars in Chapter 5.....	196
5.3 Comparisons between the Jezero and Oxia Planum sediment fans in Chapter 5.....	200
5.4 Supercam example of conglomerates at Jezero crater in Chapter 5.....	201
5.5 Satellite image of white sands erg margin in Chapter 5.....	209
5.6 Martian global database of sediment fans with HiRISE coverage in Chapter 5.....	211

List of Tables

1.1 Imaging instruments and resolutions from selected Mars orbital missions in Chapter 1.....	23
1.2 Summary of Mars surface missions in Chapter 1.....	25
2.1 Quantitative measurements of the scour-and-fill structures in Chapter 2.....	54
3.1 Sedimentary facies for units A through K in Chapter 3.....	94-96
3.2 Measurements of Apparent Average Laminae Thickness for Units A through K in Chapter 3.....	97
4.1 Summary of photogeologic units in Chapter 4.....	152
4.2 Quantitative data for sediment bodies A through G in Chapter 5.....	174
5.1 Martian sediment fans with at least two distinct stages of deposition in Chapter 5.....	203

Chapter 1: Introduction and Literature Review

1.1 Introduction

Since Mariner 9 (1971) first provided images of vast, dendritic valleys likely carved by liquid water on the surface of Mars (Figure 1.1a; e.g. Masursky, 1973; Sharp and Malin, 1975), a central goal of Mars exploration has been to understand how the Martian climate evolved from one that once supported liquid water to the cold, hyperarid desert we see on modern Mars. Over the last 50 years, a total of 26 successful missions have subsequently visited the planet, with 9 missions still active today. The ongoing ground-based and orbital-based missions continue to enable exploration of Mars in greater detail. Orbital missions produce high-resolution datasets enabling characterisation of metre-scale features across more and more of the Martian surface, while rovers systematically traverse bedrock, enabling detailed reconstructions of past depositional environments on Mars in-situ. This thesis uses high-resolution orbital and rover datasets to investigate two stratigraphic successions, at Oxia Planum and Gale crater, during a time when Mars is thought to have been transitioning towards progressively colder and more arid conditions. The results from both areas are used to consider the climatic and environmental implications for late-stage aqueous activity on Mars. This chapter provides an introduction to the geology of Mars, with the research directions for the two study areas outlined at the end of this chapter. Chapters 2-4 are results and discussion chapters; these also include detailed methodology and the discussions of the literature specific to the two main study areas. Chapter 5 provides a synthesis of the results, general conclusions, and outlines the suggested directions for future research.

1.2 A Brief History of Evidence of Surface Water on Mars

Evidence for water on Mars has long been debated since dark lineations and patches observed through telescopes were interpreted as canals and seas (e.g. Lowell, 1908; Schiaparelli, 1893). The advent of spacecraft exploration in the 1960s, however, revealed a dry, cratered world with a thin atmosphere that was not capable of supporting surface water (e.g. Kliore et al., 1965; Leighton et al., 1965). Since then, 26 successful missions have visited Mars - with 9 still operational – and have been progressively building a strong case that surface water was once widespread on the Martian surface. This section briefly summarises how each generation of missions has progressively strengthened the case for past widespread surface water on Mars for both orbiters (Table 1.1) and grounded missions (Table 1.2).

1.2.1 Orbiters

The first orbital evidence for surface water came from Mariner 9 in 1971 (Masursky, 1973), which

revealed extensive branching channels known as valley networks, widely regarded as some of the first strongest indicators that liquid water once flowed across the Martian surface (Figure 1.1a; e.g. Masursky, 1973; Sharp and Malin, 1975). The Viking orbiters that followed provided global coverage of the Martian surface at 200 – 400 m/pixel (Soffen and Snyder, 1976), showing that valleys were primarily contained to the ancient southern highlands, that their drainage networks were likely immature, and large undissected regions occurred in between individual systems leading to the hypothesis that groundwater rather than precipitation and surface runoff led to these systems (e.g. Carr, 1995; Pieri, 1980). Possible sediment fans (e.g. Cabrol et al., 1999; Ori et al., 2000) and palaeolakes defined as topographic lows with nearby fluvial inlets (e.g. Goldspiel and Squyres, 1991) were also identified using datasets from the Viking orbiters.

The recognition that surface water was once widespread on Mars advanced considerably with the arrival of high-resolution orbital datasets in the late 1990s and 2000s. The Mars Orbiter Camera (MOC; Malin et al., 1992) arrived in 2001 (Malin and Edgett, 2001) and provided the first high-resolution targeted images at 1.5–12 m/pixel, capturing the first evidence for sediment fans (Figure 1.1b; Malin and Edgett, 2003; Wood, 2006), layered sedimentary rocks (Malin and Edgett, 2000), and active gullies possibly produced by liquid water (e.g. Malin et al., 2006). After THEMIS achieved effectively global coverage of Mars by 2007 at ~100 m/pixel (e.g. Christensen et al., 2007), global investigations into the Martian surface performed at higher resolutions than before showed that valley networks were over eight times more abundant than previously recognised, with drainage characteristics consistent with precipitation-fed runoff (Hynek et al., 2010). This coverage also enabled the first global catalogue of sediment fans (Kraal et al., 2008; Moore and Howard, 2005) and palaeolakes (Fassett and Head, 2008) demonstrating that evidence for surface water was both widespread and expressed in a variety of different depositional environments.

These global catalogues of aqueous depositional environments were further expanded and refined after the arrival of the Context Camera (CTX; Malin et al., 2007) at ~5 m/pixel, achieving effectively global coverage in 2018 (Dickson et al., 2018). CTX global coverage increased the known inventory of sediment fans from a few dozen to over a thousand (e.g. Holo et al., 2021; Morgan et al., 2022; Wilson et al., 2021) and led to the identification of hundreds of open-basin or exoreic lakes (Goudge et al., 2012). Refined global mapping of valley networks with CTX datasets confirmed that most surface water flowed prior to ~3.7 Ga (Alemanno et al., 2018). Targeted images from High Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007) at 0.25 - 1 m/pixel enabled more detailed mapping of channels and sediment fans to show that fluvial activity persisted, albeit at a significantly reduced scale, into the last 2 Ga (e.g. Grant and Wilson, 2019; Kite, 2019; Wilson et al., 2021), far later than previously thought. HiRISE images also captured small amounts of possible modern aqueous activity, including active gully modification (Dundas and McEwen, 2015) and seasonally recurring slope lineae (McEwen et al., 2011), though these have later been attributed to CO₂ frost processes (Dundas et al., 2022) or aeolian processes (Dundas, 2020) without the need for liquid water.

Hyperspectral instruments aboard orbiters also provided complementary evidence for aqueous alteration of the Martian crust by directly detecting hydrated minerals. Arrival of the hyperspectral camera OMEGA (~ 0.5 km/pixel) on Mars Express in 2003 (Bibring et al., 2006) led to the first identification of hydrated silicates, specifically phyllosilicates, and hydrated sulfates (Bibring et al., 2005; Bibring et al., 2006; Poulet et al., 2005), recording a possible change in mineralogy over geologic time that may have reflected a changing climate (See Section 1.4). Subsequent higher resolution data from CRISM, arriving in 2006 (Murchie et al., 2007), resolved thousands of additional smaller exposures of phyllosilicates at 15 - 19 m/pixel showing that hydrated minerals were widespread across the highlands (Murchie et al., 2009; Mustard et al., 2008). CRISM also led to recognition of a larger diversity of minerals directly or indirectly attributed to alteration or precipitation from water across the Martian crust including carbonates, chlorides, and other hydrated silicates (Ehlmann and Edwards, 2014). CaSSIS, which arrived in 2016 and is still active today, continues to aid in identifying minerals attributed to precipitation and alteration from water across the surface of Mars (e.g. Bickel et al., 2024).

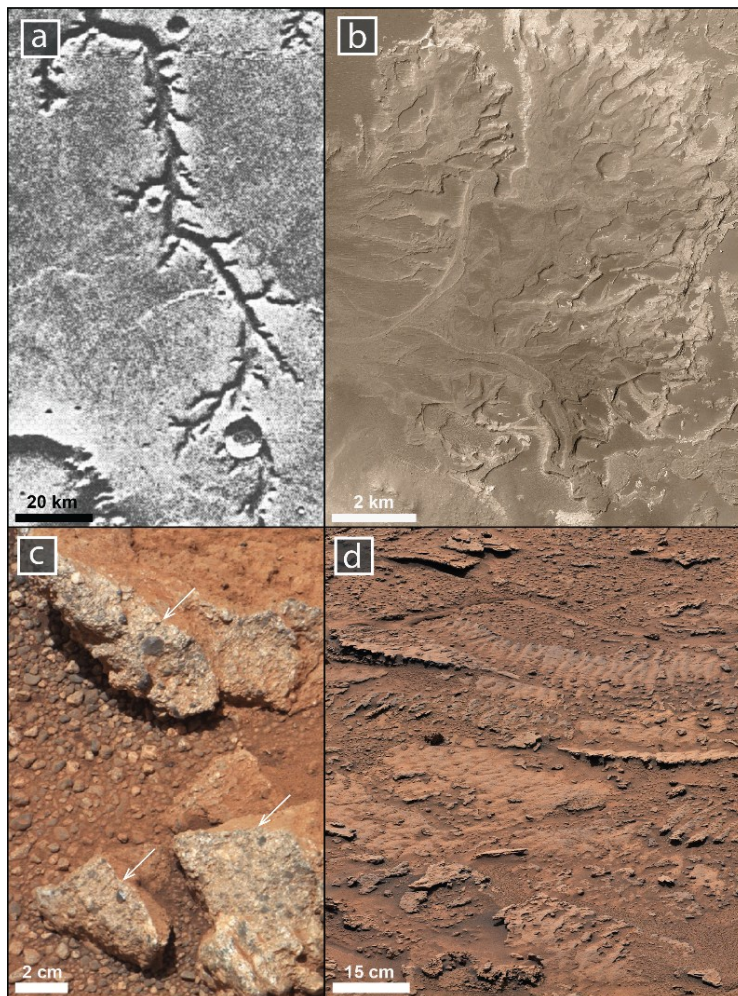


Figure 1.1: Selected evidence for water on Mars. a) Nirgal Vallis taken by Mariner 9 (from observation 06354843), one of the first pieces of evidence for past liquid surface water (e.g. Masursky, 1973; Sharp and Malin, 1975). b) The Eberswalde fan taken by the Mars Orbiter Camera in 2003 (from observation MOC2-543), the first evidence for a sediment fan on Mars (e.g. Malin and Edgett, 2003). c) A conglomerate at the landing site of Curiosity taken in 2012 (Mastcam mosaic MR_mcam00177 acquired on Sol 27, at Sol 26 end-of-drive location), the first in-situ evidence for fluvial activity (e.g. Williams et al., 2013). d) Planform symmetric ripples taken by the Curiosity rover (Mastcam mosaic MR_mcam103274 acquired on Sol 3684, at Sol 3672 end-of-drive location), the first in-situ evidence for ice-free lakes on Mars (see section 3.4.3; Mondro et al., 2025a; Mondro et al., 2025b). Image credits: NASA/JPL-Caltech/MSSS.

Instrument	Type	Resolution (m/pix)	Coverage	Mission	Years Active	Reference
Mariner 9	Visual	~1000-3000 (global), ~98 (targeted)	85%	Mariner 9	1971-1972	Masursky (1973)
Viking IS	Visual	150-300	100%	Viking IS	1975-1980	Soffen and Snyder (1976)
<i>MOLA</i>	<i>Topography</i>	<i>150 m spot; 300 m spacing</i>	<i>Global</i>	<i>Mars Global Surveyor</i>	<i>1999-2006</i>	<i>Zuber et al. (1992)</i>
MOC	Visual	2-8	Targeted	Mars Global Surveyor	1999-2006	Malin et al. (1992)
THEMIS	Visual/NIR & IR (450 to 850 nm)	~18 (vis), ~100 (IR)	Targeted	Mars Odyssey	Since 2001	Christensen et al. (2004)
HRSC	Visual	10-60	>97%	Mars Express	Since 2003	Jaumann et al. (2007)
OMEGA	Hyperspectral (350 to 5100 nm)	~300->10,000	>90%	Mars Express	Since 2003	Bibring et al. (2006)
CRISM	Hyperspectral (362 to 3920 nm)	15-19	Targeted	Mars Reconnaissance Orbiter	2006-2023	Murchie et al. (2007)
<i>CTX</i>	<i>Visual</i>	<i>~6</i>	<i>>99.5%</i>	<i>Mars Reconnaissance Orbiter</i>	<i>Since 2006</i>	<i>Malin et al. (2007)</i>
<i>HiRISE</i>	<i>Visual</i>	<i>0.25-1</i>	<i>Targeted</i>	<i>Mars Reconnaissance Orbiter</i>	<i>Since 2006</i>	<i>McEwen et al. (2007)</i>
<i>CaSSIS</i>	<i>Multispectral (four bands centred at 495 nm, 675 nm, 836 nm, and 939 nm)</i>	<i>~4.6</i>	<i>Targeted</i>	<i>ExoMars Trace Gas Orbiter</i>	<i>Since 2016</i>	<i>Thomas et al. (2017)</i>

Table 1.1: Imaging instruments, image resolution, and coverage from selected Mars orbital missions. Adapted from Palucis and Morgan (2020a) with additional missions compiled by author. See supplementary files for a complete record of all successful and failed missions to Mars. Italicised instruments are used in this thesis.

1.2.2 Landers and Rovers

Landers and rovers provide the opportunity for in situ exploration and small-scale measurements with the precision needed to confirm that surface water existed on the planet's surface. The Viking lander (1976) provided the first direct tests of Martian soils, detecting sulfates that might have been the hydrated kieserite (e.g. Clark et al., 1976), but found no direct evidence that liquid water existed on the planet. The Sojourner rover (1997) identified abundant subangular boulders, inferred to have been transported by flooding (Basilevsky et al., 1999; Rover Team, 1997), though alternative volcanic origins were also proposed (e.g. Golombek et al., 1999). It was not until the Mars Exploration Rovers, Spirit and Opportunity, landed in 2004, that the first in-situ compelling evidence for past surface water was identified. Spirit found evidence for goethite, a mineral formed in aqueous environments (Morris et al., 2006) and later sulfate minerals and opaline-A deposits produced in hydrothermal environments (Arvidson et al., 2010; Ruff et al., 2011; Squyres et al., 2008). Opportunity identified sedimentary structures in the Burns Formation attributed to deposition in aeolian environments influenced by near-surface water, including candidate festoon cross-laminations from periodic flooding and possible shrinkage cracks from wetting and drying cycles (Edgar et al., 2014; Grotzinger et al., 2006; Grotzinger et al., 2005).

More recent rover missions have provided the clearest in-situ evidence for surface water on ancient Mars. In 2012, the *Mars Science Laboratory's* rover, Curiosity, identified bedrock at the landing site which was comprised of rounded pebbles and sand grains typical of fluvial conglomerates, with the roundedness of the pebbles indicating substantial abrasion from fluvial transport (Figure 1.1c; Williams et al., 2013), confirming with in-situ evidence, that rivers once flowed across the surface of Mars. Further up the traverse, laminated mudstones were interpreted as lacustrine deposits (Grotzinger et al., 2015; Grotzinger et al., 2014), the first in-situ evidence for lakes. Continued exploration over the last 13 years has also led to the identification of other textbook examples of sedimentary structures that show evidence of water including symmetric ripples, the first in-situ evidence for ice-free lakes on Mars (Figure 1.1d; Gupta et al., 2023b; Mondro et al., 2025a). Curiosity was also the first rover to drill into bedrock, returning quantitative evidence of phyllosilicate minerals, with abundances up to 34 wt% (Bristow et al., 2021a; Thorpe et al., 2021). More recently, the Perseverance rover, landing in 2021, identified clinoforms, the first in-situ evidence for a delta on Mars (Mangold et al., 2021). Taken together, the combination of orbital and rover missions have built a consistent picture of a planet that once had a climate capable of sustaining widespread surface water.

Mission	Agency	Instruments	Years Active	Mission Goals
Viking Landers	NASA	Cameras; seismometer; biology experiment; gas chromatography; spectrometers	1976-1982	Search for biosignatures, characterise the composition of the surface.
Sojourner Rover (Mars Pathfinder)	NASA	Cameras; spectrometers	1997-1998	Investigate past water, characterise the geology and surface processes.
Spirit Rover (MER-A)	NASA	Mast & microscopic cameras; IR, Mössbauer & X-Ray spectrometers; rock abrasion tool	2004-2010	Investigate past water activity, characterise the geology and surface processes.
Opportunity Rover (MER-B)	NASA	Same as Spirit	2004-2018	Assess habitability and evidence for past water activity, characterise the geology and surface processes.
Phoenix Lander	NASA	Scoop; cameras; soil chemistry & environmental sensors	2008	Study arctic soils, search for water ice, assess habitability
Curiosity Rover (MSL)	NASA	Drill (~5 cm); mast & microscopic cameras; laser & X-ray spectrometers; sample analysis suite; radiation & environmental sensors	2012-present	Assess habitability, detect organics, characterise the geology, surface processes, and climate.
InSight Lander	NASA	Seismometer, heat probe, weather sensors, radio science	2018-2022	Study Mars's interior structure and seismicity.
Perseverance Rover (Mars 2020)	NASA	Mast & zoom cameras; Raman, X-ray & UV spectrometers; ground-penetrating radar; environmental sensors	2021-present	Assess habitability, search for biosignatures, cache samples, characterise the geology and surface processes.
Zhurong Rover (Tianwen-1)	CNSA	Navigation & multispectral cameras, ground-penetrating radar, climate sensors, magnetometer	2021-2022	Assess habitability, characterise the geology, the subsurface, the climate, and magnetic field.
<i>Rosalind Franklin Rover</i>	<i>ESA</i>	<i>Drill (2 m subsurface), analytical lab (Raman spectrometer, GC-MS, microscopy suite), cameras</i>	<i>Launch ~2028</i>	<i>Assess habitability, search for biosignatures beneath surface, characterise geology, surface processes, and climate.</i>
<i>Tianwen-3 Lander</i>	<i>CNSA</i>	<i>Drill; cameras</i>	<i>Launch ~2028</i>	<i>Sample return mission to Earth.</i>

Table 1.2: Summary of Mars surface missions, their main instruments, operational periods, and science objectives. Instrument lists are simplified to show core capabilities (e.g. cameras, spectrometers, drills, seismometers, chemistry analysers). *Italicised missions are ones which are planned for the future.* (Updated from: <https://exploration.esa.int/web/mars/-/56504-missions-to-mars>).

1.3 The Martian Climate Conundrum

Despite the growing body of evidence that liquid water once existed on the surface of Mars (Figure 1.1), liquid water - or even brine (e.g. Martínez and Renno, 2013) - is not stable under current surface conditions on Mars (Figure 1.2). Moreover, about 80 - 99% of water that was in contact with the atmosphere has been lost to space (Jakosky, 2021). This loss to space is primarily due to a combination of planetary characteristics and atmospheric loss processes that have resulted in Mars having a much thinner atmosphere than that of Earth. As the radius of Mars (~3389 km) is considerably smaller than Earth's (~6371 km), and its mass and volume are correspondingly lower, at ~0.1 and ~0.15 of Earth's, respectively (e.g. De Pater and Lissauer, 2015), there is a weaker gravitational pull and a lower escape velocity, making it easier for volatile gases to escape to space (e.g. Zahnle and Catling, 2017). In addition, unlike Earth, Mars does not currently have an active magnetic field, and the atmosphere cannot be protected from erosion by the solar wind (e.g. Dehant et al., 2007). Magnetic anomalies preserved in the oldest martian terrains suggest that if a global magnetic field ever existed, activity had ceased by ~4 Ga (Acuña et al., 1999; Solomon et al., 2005). Thus, over time, the Martian atmosphere also thinned as a result of solar wind-induced sputtering and photochemical escape (Jakosky et al., 1994), resulting in an atmospheric pressure averaging ~6.1 mbar (e.g. De Pater and Lissauer, 2015; Kliore et al., 1965). This low atmospheric pressure severely restricts the range of temperatures at which liquid water can remain stable on the surface compared to Earth (Figure 1.2). In addition, the thin atmosphere is inefficient at retaining heat. The mean global temperature of Mars is ~215 K (-52°C), with an average diurnal range of ~60 K (De Pater and Lissauer, 2015) meaning that in most places liquid water cannot stably exist. Thus, a climate paradox emerges: how did the planet sustain a climate that permitted the formation of the abundant aqueous depositional and erosional features observed on ancient Mars?

This climate paradox is even more complicated when looking back at the geologic history of Mars, under a faint young sun at luminosities ~25% weaker than the modern day (e.g. Gough, 1982). Dense atmospheres with high atmospheric pressures would be required to produce a climate capable of sustained surface liquid water (Figure 1.2). However, one-dimensional climate models (e.g. Kasting, 1991; Ramirez et al., 2014) and three-dimensional global simulations of the early Martian climate (Forget et al., 2013; Mischna et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2015) have so far been unsuccessful in producing long-lived conditions capable of sustaining surface temperatures above 0°C. Moreover, sparse carbonate-rich deposits identified from orbit have been argued to be insufficient to account for a CO₂ dense atmosphere in the geologic past (e.g. Niles et al., 2013). However, recent research suggests that sufficient CO₂ may have been sequestered in other deposits like sulfate-bearing rocks (e.g. Tutolo et al., 2025) and lost to space (Jakosky, 2019). Nonetheless, most climate models typically predict cold and icy conditions even with high atmospheric densities using CO₂ and H₂O vapour (Forget et al., 2013; Wordsworth et al., 2013), with an increased planetary albedo resulting from hypothesized ice sheets making it even more difficult to transiently warm the surface

(Ramirez, 2017).

This climate paradox leads to two different models on Mars. The “warm and wet” climate hypothesis supported by the orbitally deduced evidence for widespread precipitation and surface runoff (e.g. Craddock and Howard, 2002; Moore et al., 2003). Alternatively, there is a “cold and icy” hypothesis which is supported by the climate models (Forget et al., 2013; Wordsworth et al., 2013), suggesting that Mars underwent punctuated, episodic warming that could produce some of the large-scale fluvial features before returning to ice-house conditions. These periods of episodic warming could have been caused by volcanism (Halevy and Head III, 2014), large impacts (Segura et al., 2012), varying eccentricities or obliquities (e.g. Kite and Conway, 2024), or high-altitude water ice clouds (e.g. Kite et al., 2021). However, recent modelling suggests these conditions could not have lasted for more than 10^5 - 10^7 years (Kamada et al., 2020; Kite and Conway, 2024; Ramirez et al., 2020) and there is a scarcity of evidence for glacial and ice-sheet features in the low- to mid-latitudes that would be necessary for an ice-house global climate (e.g. Grotzinger et al., 2015).

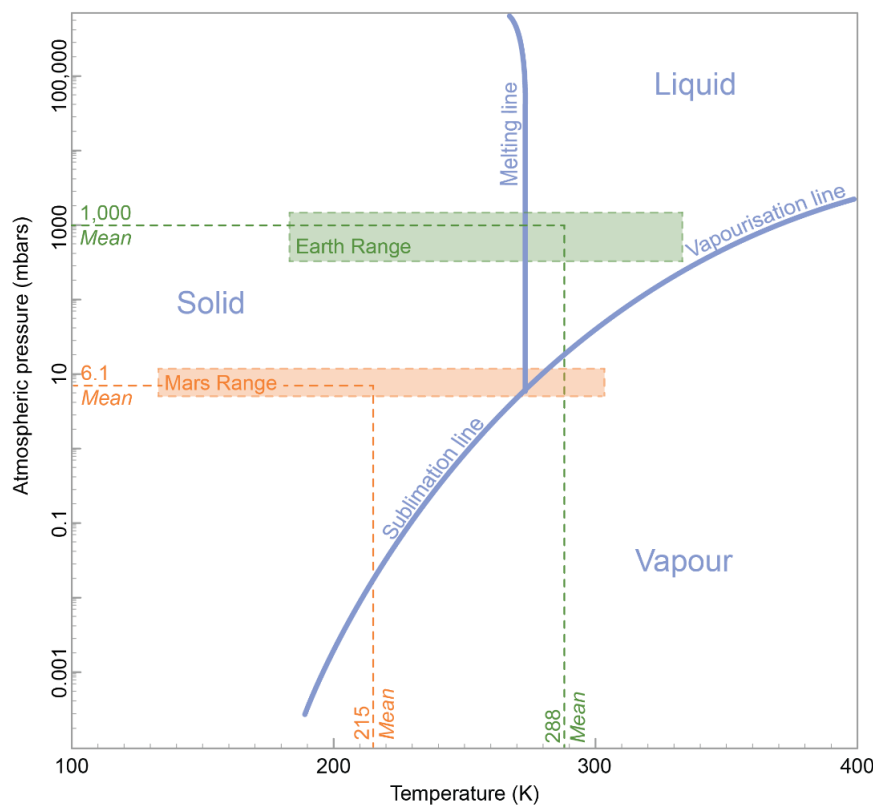


Figure 1.2: Phase diagram for the triple point of water set against the range of surface temperatures and pressures on Mars and Earth. Note the limited range of Mars’s conditions where water can be in liquid form on the surface. Adapted from Verseux (2018).

1.4 Geologic history of Mars

Orbital, landed, and rover datasets over the last five decades have shown that the planet once hosted widespread aqueous activity, indicating that the climate was once warm enough to support surface water. However, this long-term climate evolution cannot be fully reconstructed without reference to the planet's broader geologic history. In addition to aqueous activity, Mars was also shaped by a number of surface processes including impact cratering, volcanism, tectonics, and wind. The following sections outline the major physiographic controls on Mars, introduce the Martian geologic timescale, and summarise the key events of each geologic period.

1.4.1 Distinct Physiographic Controls on Mars

Mars's physiography is defined by several surface processes distinct to Earth, each of which has strongly influenced the planet's geological history. The dominant surface-shaping process throughout Martian time has been impact cratering, ranging from sub-10 m wide craters to the ~3,300 km diameter basin of Utopia Planitia. Most impact craters smaller than ~1 km in diameter are likely secondaries, formed by ejecta from larger impacts rather than by primary projectiles (e.g. McEwen et al., 2005). Volcanism on Mars also differs significantly to Earth. In the absence of plate tectonics, volcanism resulting from mantle sources is static on Mars (e.g. Sleep, 1994), resulting in the long-lived, large volcanic provinces such as Tharsis and Elysium that reach up to hundreds of kilometres in diameter. In particular, this form of volcanism led to the formation of Olympus Mons, which, at ~26 km high, is the tallest volcano in the solar system. Unlike Earth, tectonic activity on Mars is also not driven by plate tectonics. Instead, formation of Tharsis likely generated immense tectonic stresses that produced crustal shortening and thrust-fault landforms across Mars (e.g. Watters, 1993; Woodley et al., 2024), and may have initiated the >4000 km long, 200 km wide, and up to 10 km deep Valles Marineris canyon system, later enlarged by erosion and weathering (e.g. Andrews-Hanna, 2012). Finally, one of the most notable and oldest features on the Martian surface is the crustal dichotomy, a pronounced elevation contrast between the heavily cratered, thick-crustal southern highlands and the lower, thinner-crustal northern lowlands buried beneath regolith and lithified rock (e.g. Hartmann, 1973; Knapmeyer-Endrun et al., 2021; Neumann et al., 2004; Tanaka et al., 2014). Though more recent findings from the seismometer aboard the *Insight* lander suggest that this crustal thickness is not as significant as previously suggested from datasets derived by MOLA (Wieczorek et al., 2022). This dichotomy potentially formed from giant impact events (e.g. Ballantyne et al., 2023; Citron et al., 2018; Marinova et al., 2011), mantle convection (e.g. Roberts and Zhong, 2006), or even a hybrid model with a positive feedback effect (Bonnet Gibet et al., 2022; Roberts and Arkani-Hamed, 2017), leading to excavation and thinning of the northern crust from mantle upwelling.

1.4.2 Introduction to Martian Geologic Time

The Martian geologic timescale is based on relative ages from superposition relationships and size-frequency distributions of impact craters. This timescale is divided into three main periods - the Noachian, Hesperian, and Amazonian - which are further subdivided into eight epochs (Figure 1.4). The period from planetary accretion until the start of the Noachian is regarded as the Pre-Noachian and is not preserved on the surface but had an influence on the planet's current topography. Another suggested geologic timescale uses Mars' surface mineralogy, which divided the Martian terrain into three periods based on different dominant mineral species detected from orbit: a phyllosilicate-forming period (Phyllocian), a sulfate-forming period (Theiikian), and an iron oxide-forming period (Siderikan) (Figure 1.3; Bibring et al., 2006). This thesis will use the geologic timescale based on crater density.

Absolute ages on Mars are hard to constrain. In situ measurements of absolute ages of bedrock on extraterrestrial environments are currently not available - with the exception of a radiogenic K-Ar age of 4.21 ± 0.35 billion years taken from a mudstone in Gale crater by the Curiosity rover (Farley et al., 2014). Instead, relative ages are often obtained based on cross-cutting relationships and the size-frequency distribution of impact craters across a surface, as older surfaces have had more time to accumulate impacts and larger impacts were more common early on in Mars' history (e.g. Hartmann, 1966; Hartmann and Neukum, 2001). Though this method works best on surfaces larger than 100,000 km², as smaller surfaces can artificially exclude statistically significant large craters, affecting the result (Warner et al., 2015). For Mars, absolute ages are calculated by comparing the martian impactor flux to the lunar impactor flux adjusting for Mars' size and gravity, as lunar rock samples have been radiogenically dated on Earth (e.g. Hartmann, 2005). As this process relies on a number of assumptions, multiple different crater production and chronology functions have been created which provide different absolute ages for the same terrain (Figure 1.3; e.g. Fassett, 2016). Due to the small sizes of the given study areas, this thesis does not often give absolute ages of a surface but instead refers to their geologic time period and epoch. When given, this work uses the chronology of Michael (2013).

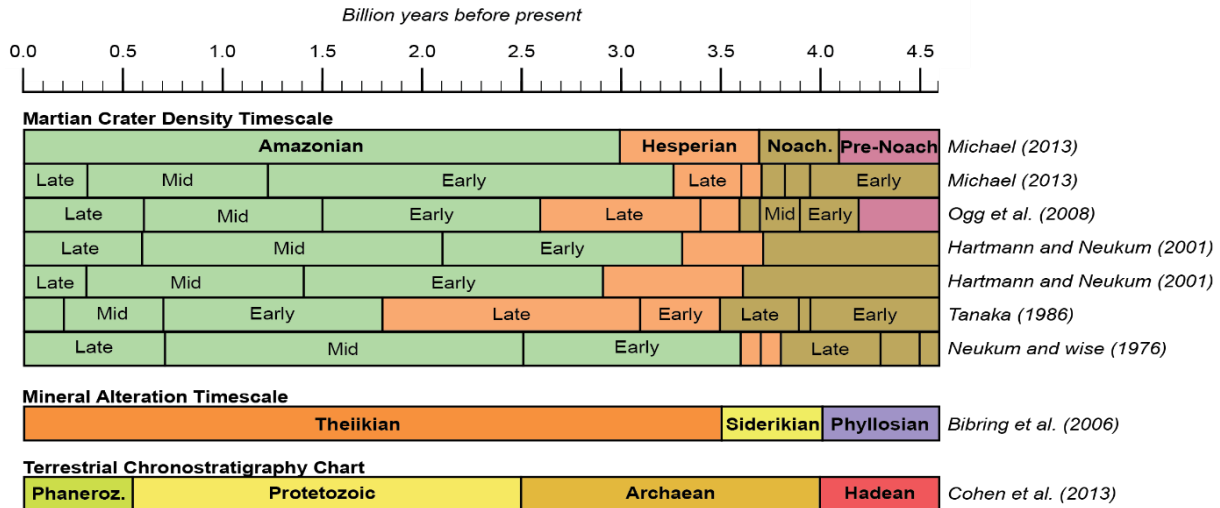


Figure 1.3: Variety of Martian geologic timescales based on crater density and surface mineralogy compared to the terrestrial chronostratigraphy chart. This thesis uses Michael (2013). Adapted from Davis (2017).

1.4.3 Pre-Noachian Mars

Pre-Noachian Mars commenced from the accretion of the planet and is generally taken to end at the ~1700 km-diameter Hellas impact basin at ~4.1 Ga (Figure 1.3; Figure 1.4a; e.g. Carr and Head, 2010). Much of the geologic record of this interval has been erased by subsequent erosion and high impact rates (Figure 1.5; e.g. Frey, 2008). Magnetic anomalies in the ancient crust suggest that the planet has a strong magnetic field for the majority of this time period (e.g. Acuña et al., 1999; Solomon et al., 2005). A number of large impact basins also formed during this time period, identified from quasi-semi-circular depressions in MOLA and derived crustal thickness datasets (e.g. Frey, 2008; Frey, 2006; Nimmo and Tanaka, 2005), including the largest and generally most well-accepted impact basin, Utopia (Figure 1.4a; Frey, 2008; Frey, 2006; Neumann et al., 2004; Nimmo and Tanaka, 2005) and terminated with the formation of Hellas (e.g. Carr and Head, 2010). The formation of the crustal dichotomy likely commenced during the Pre-Noachian, evidenced by superimposed large impact basins (Figure 1.5; e.g. Irwin III and Watters, 2010; Neumann et al., 2004) and geochemical arguments (e.g. Bottke and Andrews-Hanna, 2017; Brasser and Mojzsis, 2017). Some researchers suggest that the dichotomy continued developing through the Pre-Noachian, ceasing in the early Hesperian (e.g. Bonnet Gibet et al., 2022; Citron et al., 2018).

1.4.4 Noachian Mars

The Noachian geologic period is generally associated with the formation of the southern highlands (Figure 1.4a; Tanaka et al., 2014). Much of the primary crust from this period is thought to consist of volcanic rocks, either emplaced directly or reworked by impacts, and altered by hydrothermal processes (Carr and Head, 2010), similar to the bedrock observed by the Spirit rover (e.g. Squyres et al., 2006). Stratified, clay-bearing deposits are also abundant, with ~88% of exposures occurring in the Noachian-

aged highlands (e.g. Ehlmann and Edwards, 2014; Tanaka et al., 2014; Ye and Michalski, 2022). The period coincided with the late heavy bombardment, resulting in a high rate of impact cratering (Figure 1.5; Bottke and Norman, 2017; Frey, 2008). Volcanism was also active, with the construction of the large volcanic province, Tharsis Rise (Figure 1.4a), beginning in the Mid-Noachian (Figure 1.5; Carr and Head, 2010; Fassett and Head, 2011). Morphological evidence suggests that eruptions during this period were comparatively more explosive than later stages of Mars's history, attributed to higher volatile content in magma (e.g. Brož et al., 2021; Kremer et al., 2019; Robbins et al., 2011), which in turn may have contributed larger volumes of volcanic gases to the atmosphere (e.g. Jakosky, 2019).

The Noachian is thought to have two significant periods of aqueous activity between long haituses (Kite and Conway, 2024). The earliest record of aqueous activity occurred during the early- to mid-Noachian. During this time, there is the first evidence for fluvial activity (Figure 1.5; e.g. Balme et al., 2020; Davis et al., 2023; Kite and Conway, 2024; Skinner et al., 2021). Flat-bottomed craters from this time period are also attributed to infilling from water or ice-slurry (El-Maarry et al., 2013; Forsberg-Taylor et al., 2004; Kite and Conway, 2024), and the bimodal distribution of crater infill has been interpreted as evidence for streams or debris flows (Forsberg-Taylor et al., 2004). Widespread phyllosilicates and Fe-rich chlorites (Figure 1.5; Mustard et al., 2008), minerals formed through aqueous alteration of basalts (e.g. Zolotov and Mironenko, 2007), further indicate aqueous activity was extensive and may have begun even prior to the onset of the Noachian (Ehlmann and Edwards, 2014; Tanaka et al., 2014; Ye and Michalski, 2022).

The second period of aqueous activity occurred in the late Noachian just prior to the Noachian-Hesperian boundary (e.g. Kite and Conway, 2024), coinciding with a climatic optimum when erosion rates peaked (e.g. Howard et al., 2005; Irwin et al., 2005) and the majority of the valley networks formed (Figure 1.4b; Fassett and Head, 2008). These networks are typically several to tens of kilometres wide and hundreds of kilometres long (e.g. Williams and Phillips, 2001), with formation timescales exceeding 10,000 years (e.g. Orofino et al., 2018). The dendritic appearance of the channels implies that precipitation might have been a driving component of overland flow (e.g. Davis et al., 2016; Fassett and Head, 2008; Hynek et al., 2010) and mineralogical evidence may support this interpretation, as stratified Al-clays overlying Fe/Mg-clays suggest that aqueous fluids leached the Martian regolith, possibly as a result of rainfall or snowmelt (e.g. Loizeau et al., 2018). Crater palaeolake basins (e.g. Boatwright and Head, 2022) and some sediment fans (Figure 1.5; e.g. Holo et al., 2021; Wilson et al., 2021) are also attributed to this climatic optimum. The possibility of a northern ocean has likewise been proposed during the Noachian/Hesperian boundary, as many of these sediment fans are putative deltas that form at the Martian dichotomy (Di Achille and Hynek, 2010; Fawdon et al., 2018; Rodriguez et al., 2016; Rodriguez et al., 2022). Yet this hypothesis remains difficult to substantiate (e.g. Sholes et al., 2021).

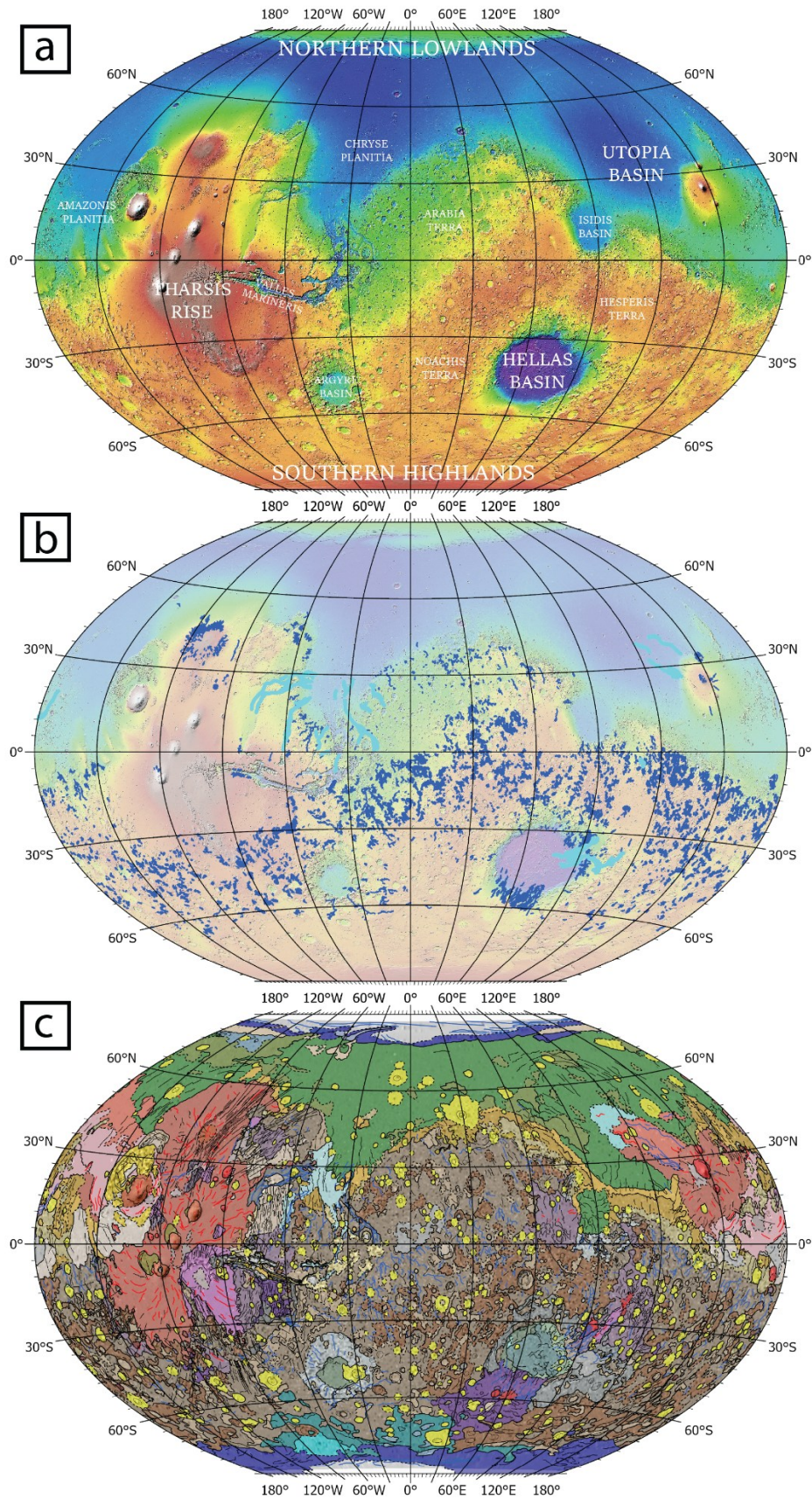


Figure 1.4: a) Gridded MOLA topographic map of Mars with major regions labelled. b) Valley networks (dark blue) and outflow channels (light blue) as mapped by Alemanno et al. (2018). c) Geological map of Mars from Tanaka et al. (2014) showing the Noachian highland units (brown), major volcanic provinces (reds and purples), Amazonian lowland units (green) and impact craters (yellow).

1.4.5 Hesperian Mars

Significant volcanism persisted through the Hesperian, leading to large flood basalt deposits that may have resurfaced 30% of the planet (e.g. Carr and Head, 2010), considered as a possible source of short-lived episodic climate-changing volcanism (e.g. Halevy and Head III, 2014). Continued magmatic activity at Tharsis caused substantial loading and deformation of the elastic lithosphere and the migration of the location of the rotational pole of Mars (e.g. Bouley et al., 2016; Citron et al., 2018; Perron et al., 2007), potentially off-setting the crustal dichotomy from an earlier equatorial alignment (Bouley et al., 2016). At the same time, Mars lost substantial CO₂ and H₂O to space (Haberle et al., 2017), impact rates dropped to near-modern levels, and erosion rates declined relative to the Noachian but remained above Amazonian values (Golombek and Bridges, 2000). Aqueous activity during the Hesperian is not well constrained. However, there is a growing body of evidence for two significant periods of aqueous activity: a period of waning aqueous conditions from the climate optimum and a late Hesperian to early Amazonian resurgence in aqueous activity.

The Hesperian experienced a protracted period of waning aqueous activity from the climate optimum at the Noachian/Hesperian boundary. Crater chronology indicates that formation of large, integrated valley networks declined in the early Hesperian and then ceased by ~3.4 Ga along with widespread phyllosilicate alteration (e.g. Fassett and Head, 2008; Hynek et al., 2010). However, fluvial activity continued albeit with a more ‘patchy’ record (Figure 1.5; e.g. Kite and Conway, 2024; Wilson et al., 2021). During this time, the main morphology of fluvial features appeared to change to ones characterised by shorter landforms (e.g. Kite and Conway, 2024). This period was dominated by shorter, broader valleys (e.g. Hobley et al., 2014; Wilson et al., 2016), candidate crater lakes with small outlet breaches (e.g. Warren et al., 2021; Wilson et al., 2016), and <30% of the delta and alluvial fan record (e.g. Morgan et al., 2022; Wilson et al., 2021). There is also some evidence that precipitation-fed runoff occurred (e.g. Kite, 2019; Lapôtre and Ielpi, 2020), though cryospheric melting events have also been proposed (e.g. Hauber et al., 2013). Some fans reached ~1 km in thickness, recording at least 10²–10⁶ years of flow (e.g. Irwin et al., 2015; Stucky de Quay et al., 2019). Such records of fluvial activity persisted throughout the Hesperian and into the early Amazonian (e.g. Kite et al., 2017; Morgan et al., 2022; Wilson et al., 2021). Though, declining sizes of crater lake catchment areas (e.g. Kite and Noblet, 2022) and increased sporadity of the fluvial fan record (e.g. Wilson et al., 2021) suggest an increasing amount of aridity throughout the Hesperian and into the Amazonian (e.g. Kite and Conway, 2024; Kite and Noblet, 2022; Wilson et al., 2021).

The protracted waning from the Noachian/Hesperian boundary also coincides with a change in mineralogy. Hesperian geologic units are characterised by extensive sulfate-bearing layered sedimentary rocks reaching up to ~5 km in thickness (Figure 1.5; e.g. Bibring et al., 2006; Ehlmann and Edwards, 2014; Mangold et al., 2008; Murchie et al., 2009), sometimes overlying clay-rich Noachian units (e.g. Arvidson et al., 2005; Gendrin et al., 2005; Wiseman et al., 2010). The sulfate-rich

mineralogy indicates a water-related origin. Proposed origins of this global change in mineralogy to sulfate-bearing terrain include concentration of pre-existing salts through evaporation of groundwater during the aridification of Mars (e.g. Andrews-Hanna et al., 2007; Andrews-Hanna et al., 2010; Bibring et al., 2006), an anomalously high supply of acidic sulfates by Hesperian volcanism (e.g. Bibring et al., 2006), and/or preferential remobilisation of an earlier Noachian sulfate reservoir that was subsequently depleted (e.g. Zolotov and Mironenko, 2016). Sulfate outcrops have also been observed to coincide with areas of predicted groundwater upwelling, such as Terra Sirenum, Meridiani Planum, Aram Chaos, and Gale crater (e.g. Andrews-Hanna et al., 2010).

The late Hesperian to early Amazonian witnessed a second major period of aqueous activity. This was most dramatically expressed in the formation of the outflow canyons that flowed into the northern lowlands, the largest fluvial features in the solar system (Figure 1.4b; Figure 1.5; e.g. Harrison and Grimm, 2008; Palucis and Morgan, 2020b). Unlike valley networks, they lack tributaries (e.g. Baker, 1978; Carr, 1987), and begin abruptly, consistent with subsurface reservoirs instead of from precipitation and surface run-off (e.g. Palucis and Morgan, 2020b). Their morphologies such as their anastomosing patterns, tear-drop shaped islands, midchannel bars, high width to depth ratios, suggest catastrophic flooding (e.g. Baker, 1978; Baker, 2002). Associated chaos and pitted terrains indicate that flooding might be from massive ground-ice melting (e.g. Carr, 1996; Clifford and Parker, 2001). Terraces within channels could suggest multiple flooding events (e.g. Larsen and Lamb, 2016), though these could be the result of flow instabilities during a single event (e.g. Finnegan and Dietrich, 2011; Lamb and Fongstad, 2010). These events may have also led to the formation of short-lived Hesperian northern oceans (e.g. Rodriguez et al., 2015; Rodriguez et al., 2022), though these hypothesised oceans are difficult to substantiate. During the late Hesperian, hydrated sulfates continued to form (Milliken et al., 2008), as did palaeolakes (Wilson et al., 2021) and sediment fans (Figure 1.5; e.g. Morgan et al., 2022; Wilson et al., 2021). During this time, embedded impact craters have suggested a lower bound of 20 Myr to even 100 Myr durations for alluvial fan aggradation (e.g. Kite et al., 2017). However, there is widespread recognition that a better handle on the timing and duration of surface aqueous conditions post-Noachian is needed to meaningfully advance understanding of the martian climate evolution (e.g. Kite and Conway, 2024).

1.4.6 Amazonian Mars

During the Amazonian, volcanism became increasingly restricted to Tharsis and Elysium, with plains volcanism active until the last tens of millions of years (Hauber et al., 2011; Hauber et al., 2013; Werner, 2009). Surface evolution was dominated by impact cratering, aeolian processes, and ground-ice processes as water became largely sequestered in the cryosphere and polar caps (e.g. Balme et al., 2013). Aqueous activity, though limited, persisted. Alluvial fans continued to form, albeit smaller and more localised than their Hesperian counterparts, with some linked to snowmelt-driven runoff and dated to the Late Amazonian (Goddard et al., 2014; Kite, 2019; Palucis and Morgan, 2020b; Williams and

Malin, 2008). Though the formation of large sedimentary fans apparently ceased over 10^8 years ago (e.g. Kite and Conway, 2024). Sulfate formation also extended into the late amazonian, though most outcrops predate ~ 2.5 Ga (e.g. Kite and Conway, 2024; Szykiewicz and Bishop, 2021), minor wetting produced chemical alteration in the form of rinds, veneers, and crusts (Knoll et al., 2008). Today, water is largely confined to ice, yet transient liquid phases may still persist (e.g. Kite and Conway, 2024).

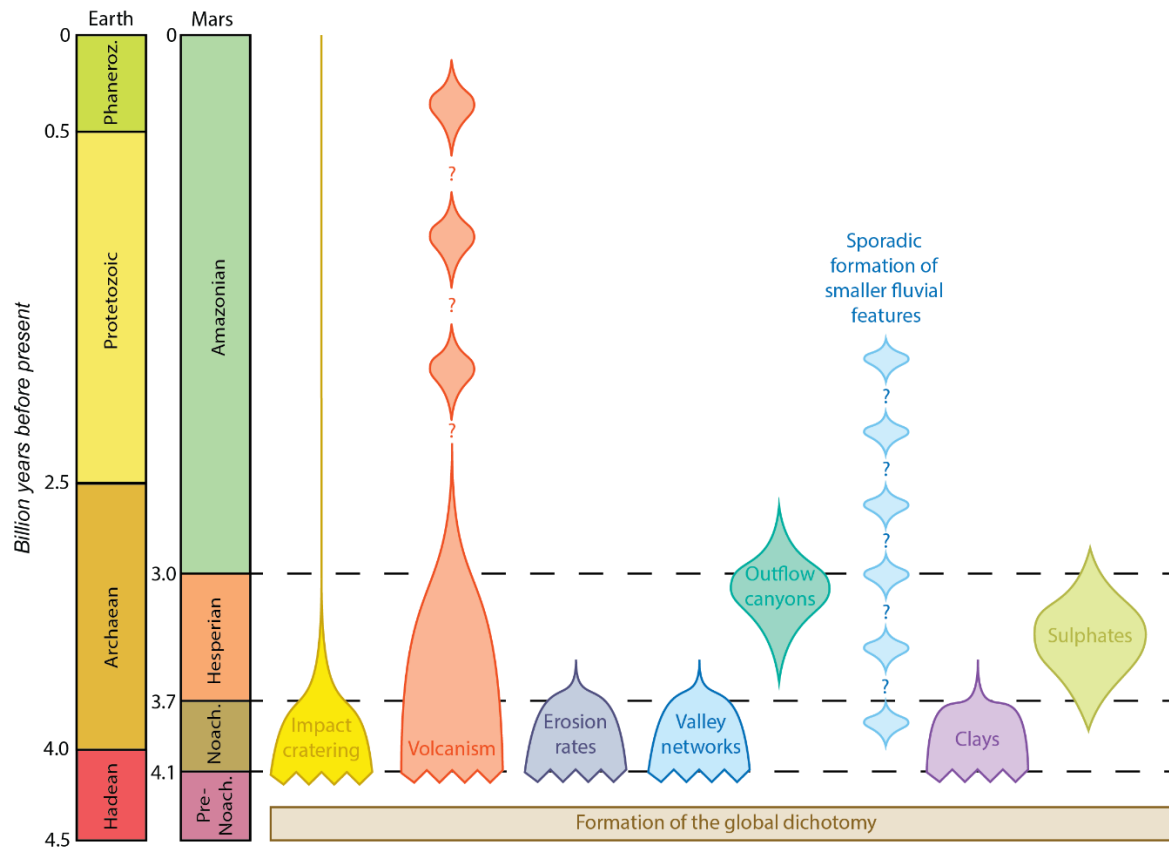


Figure 1.5: A schematic of the geologic evolution of Mars compared to the Earth's geologic timescale, showing the relative timing of major processes and events. Adapted from McNeil (2023), with additional information taken from Palucis and Morgan (2020) and Kite and Conway (2024).

1.5 Open Questions Regarding Martian Climate Change

Prior to the arrival of targeted ~ 25 cm/pixel satellite images of the surface of Mars (Section 1.1), the general consensus had been that widespread fluvial activity was restricted to the Late Noachian and Early Hesperian Periods (Carr, 2012; Carr and Head, 2010), during a fluvial optimum that permitted intensified erosion (Howard et al., 2005; Irwin et al., 2005). More recent research, however (Sections 1.4.5–1.4.6), have shown that aqueous activity persisted into the Hesperian and even the Amazonian, raising significant questions about the traditional “drying out” model. The predominance of sulfates during the Hesperian (e.g. Bibring et al., 2006), the transition into shorter fluvial landforms (e.g. Kite and Conway, 2024), and the patchy fluvial record (e.g. Kite and Noblet, 2022; Wilson et al., 2021) suggest a climate trend to one that was progressively more arid or alternatively toward briefer wet

events (e.g. Kite and Conway, 2024; Kite and Noblet, 2022). Recent research has suggested that Mars did not undergo a single wet-to-dry transition, and may have, instead, experienced multiple environmental transitions (e.g. Kite and Conway, 2024). However, the nature of this wet-to-dry transition is difficult to constrain globally given the fragmentary record of aqueous activity (e.g. Wilson et al., 2021) recorded on the surface.

Uncertainty also persists over whether early Mars was fundamentally “warm and wet” or “cold and icy” or somewhere in between (Section 1.2). In the former case, progressive loss of the water inventory and a declining groundwater table could explain aridification (e.g. Andrews-Hanna and Lewis, 2011; Jakosky, 2021; Salese et al., 2019). In the latter, changes in the greenhouse effect may have driven progressively colder surface conditions over time, increasingly too cold for liquid water (e.g. Kite et al., 2022). While this thesis cannot claim to resolve the global climate evolution of Mars, the two selected study areas - the basal section of the layered sulfates at Gale crater and the sediment fan in Oxia Planum - formed during the interval when fluvial activity began to wane after the peak of the climate optimum during the Noachian/Hesperian boundary (e.g. Fawdon et al., 2022; Thomson et al., 2011), providing an opportunity to investigate depositional environments recorded during that time.

1.6 Research Questions and Objectives

This thesis provides a detailed study into the sedimentary geology and geomorphology of two study areas on Mars, both of which have been proposed to record regional climate during and after the Noachian/Hesperian boundary. The overarching objective is to evaluate how the sedimentary environments evolved through time and the implications for the regional climate. The first study area, the basal section of the layered sulfate succession in Gale crater (Chapters 2 and 3), was explored in-situ by the Curiosity rover during the course of this PhD. The layered sulfate-bearing stratigraphy is part of a regional mineralogical transition at Gale crater identified from orbit (Fraeman et al., 2016; Sheppard et al., 2021b), the first clay-sulfate transition explored on the ground. The stratigraphic interval explored in this thesis contributes to the broader understanding of the changing sedimentary environments recorded in the clay-sulfate transition, represents the first encounter of the layered sulfate-bearing stratigraphy by Curiosity, and provides the first detailed investigation of the facies and sedimentary architectures. Using a combination of rover and orbital datasets, I (1) document the facies and sedimentary architecture of this basal section of the layered-sulfates, (2) reconstruct their depositional environments, (3) and discuss their implications for the regional climate in Gale crater at the time of their formation.

The second study area, the sediment fan at Oxia Planum (Chapter 4), lies within one of the proposed landing ellipses of the upcoming Rosalind Franklin rover and has an extensive regional catchment originating in Mars’s southern highlands (e.g. Fawdon et al., 2022). As the sediment fan was likely emplaced during the Noachian/Hesperian boundary prior to the emplacement of Kilhampton crater in the early Noachian and there is evidence for episodic activity (e.g. Fawdon et al., 2022; Quantin-Nataf

et al., 2021), this sediment fan represents an opportunity to investigate the timing and longevity of aqueous activity in the Oxia Planum basin and the regional climate during the end of the climate optimum. Investigating the sediment fan will also aid in the science goals for the upcoming mission (Vago et al., 2017). This chapter uses high-resolution orbital datasets to (1) investigate the geomorphology and photogeology of the sediment fan, (2) reconstruct the timing and longevity of aqueous activity in the Oxia Planum basin, and (3) assess the implications for the regional climate.

These two study areas were selected as the basis of study for this thesis because of (1) their similar geologic time period, (2) they are regional representative examples with excellent geological records, (3) the abundance of datasets including the opportunity to integrate orbital and in situ observations at Gale crater, (4) their relevance to ongoing and upcoming missions, and (5) their importance for understanding how climate change shaped sedimentary environments and influenced habitability during late-stage aqueous activity on Mars.

1.7 Statement of Originality and Contribution

The work presented in this thesis was completed during my PhD at Imperial College London. Everything included in this thesis is entirely my own, except where clearly stated or referenced. An abridged version of Chapter 2 has been published in a peer-reviewed journal. A large part of Chapter 3 contains work currently in preparation for publication in a peer-reviewed journal, in collaboration with members of the Curiosity Rover Team: Sanjeev Gupta, Aster Cowart, Lauren Edgar, William Rapin, William Dietrich, Edwin Kite, Steven Banham, Joel Davis, Dave Rubin, John Grotzinger, Claire Mondro, William Farrand, Alexander Bryk, Rachel Sheppard, Linda Kah, Gwénaél Caravaca, Stéphane Le Mouélic, Catherine Weitz, Aileen Yingst, Deidra Fey, Emma Harris, Jeffrey Johnson, Patrick Gasda, Abigail Fraeman, and Ashwin Vasavada. All results presented in this thesis were interpreted jointly by me and my supervisors Sanjeev Gupta, Steven Banham, Joel Davis, and Peter Fawdon.

1.8 Publications

First author papers:

- **A. L. Roberts**, S. Gupta, S. Banham, A. Cowart, L. Edgar, W. Rapin, W. Dietrich, J. Davis, E. Kite, G. Caravaca, C. Mondro, P. Gasda, J. Johnson, S. Le Mouélic, D. Fey, A. Bryk, E. Harris, A. A. Fraeman, A. Vasavada. (2025). Paleo-Scours Within The Layered Sulfate-Bearing Unit At Gale Crater, Mars: Evidence For Intense Wind Erosion. *Journal of Geophysical Research: Planets*, **130**(5). <https://doi.org/10.1029/2024JE008680>

Contributed papers:

- S. G. Banham, **A. L. Roberts**, S. Gupta, J. M. Davis, L. M. Thompson, D. M. Rubin, G. Paar, K. L. Siebach, W. E. Dietrich, A. A. Fraeman, A. R. Vasavada. (2024). Ice? Salt? Pressure?

Sediment deformation structures as evidence of late-stage shallow groundwater in Gale crater, Mars. *Geology*, **52**(7). <https://doi.org/10.1130/G51849.1>

- B. M. Tutolo, E. M. Hausrath, E. S. Kite, E. B. Rampe, T. F. Bristow, R. T. Downs, A. Treiman, T. S. Peretyazhko, M. T. Thorpe, J. P. Grotzinger, **A. L. Roberts**, et. al. (2025). In situ evidence of an active carbon cycle on ancient Mars. *Science*, **388**(6744). <https://doi.org/10.1126/science.ado9966>
- D. Zhou, M. L. Turner, W. Rapin, J. Schieber, **A. L. Roberts**, et al. (2025). How Does Topography Affect Abrasive Wind Flow on Mars? Recently Observed Shifts in Ventifact Orientation at Gale crater. *Icarus*, **437**. <https://doi.org/10.1016/j.icarus.2025.116605>
- C. Mondro, J. Grotzinger, C. Fedo, M. Lamb, S. Gupta, W. Dietrich, S. Banham, C. Weitz, P. Gasda, D. Rubin, A. Cowart, K. Lewis, **A. L. Roberts**, A. B. Bryk, R. M. E. Williams, W. Rapin, C. Seeger, E. Kite, et. al. (2025). Depositional environment of the Amapari Marker Band: rising water levels formed kilometre-scale lake in Gale Crater, Mars. *Journal of Geophysical Research: Planets*, **130**(3). <https://doi.org/10.1029/2024JE008606>
- W. Farrand, A. Eng, A. Trussell, J. Johnson, J. Bell III, S. Banham, **A. L. Roberts**, L. Thompson, E. Rampe, C. Hargrove, C. Fedo, A. Fraeman, C. Weitz. (2025). Multispectral Properties of Rocks in Marker Band Valley, and Evidence for an Alteration Unit Below the Amapari Marker Band at Gale Crater, Mars. *Journal of Geophysical Research: Planets*, **130**(4). <https://doi.org/10.1029/2024JE008606>
- J. M. Davis, S. Gupta, P. M. Grindrod, A. N. Rudolph, S. W. Purdy, J. A. Grant, R. Williams, E. S. Kite, **A. L. Roberts**, E. R. Harris, G. Paar. (2025). Late-stage aqueous activity at Gale crater, Mars, recorded by sediment fans eroded from Aeolis Mons. *Journal of Geophysical Research: Planets*, **130**(3). <https://doi.org/10.1029/2024JE008808>
- A. Rudolph, B. Horgan, K. Bennett, C. Weitz R. Sheppard, S. G. Banham, A. B. Bryk, E. Kite, **A. L. Roberts**, L. Scuderi. (2024). An Orbital Comparison of a Late Mantling Unit on Aeolis Mons With Other Erosion-Resistant Strata Explored by MSL in Gale Crater, Mars. *Journal of Geophysical Research: Planets*, **129**(8). <https://doi.org/10.1029/2023JE008242>
- E. Harris, J. Davis, P. Grindrod, P. Fawdon, **A. L. Roberts**. (2024). A Low albedo, Thin, Resistant Unit in Oxia Planum, Mars: Evidence for a Palaeo-Ashfall and Late-stage Groundwater Activity at the ExoMars Rover Landing Site. *Journal of Geophysical Research: Planets*. **129**(11). <https://doi.org/10.1029/2024JE008527>

First author conference abstracts:

- **A. L. Roberts**, S. Gupta, P. Fawdon, S. Banham, J. Davis, E. Harris. (2024). Reconstructing Depositional Environments of the Sediment Fan in Oxia Planum, Mars. Europlanet Science Congress.

- **A. L. Roberts**, S. Gupta, A. Cowart, L. Edgar, W. E. Dietrich, W. Rapin, E. S. Kite, S. G. Banham, J. M. Davis, C. A. Mondro, W. Farrand, A. B. Bryk, T. Kubacki, N. Moore, R. Y. Sheppard, L. C. Kah, G. Caravaca, A. Fraeman, A. R. Vasavada. (2024). Depositional processes and environments of the Layered Sulfate-bearing unit, Gale crater, Mars. Tenth International Conference on Mars, Pasadena, CA.
- **A. L. Roberts**, S. Gupta, W. E. Dietrich, L. A. Edgar, W. Rapin, S. G. Banham, J. M. Davis, E. S. Kite, A. Cowart, A. B. Bryk, L. C. Kah, T. Kubacki, N. Moore, P. J. Gasda, J. R. Johnson, G. Caravaca, A. Fraeman, A. Vasavada. (2023). Observations of aeolian palaeo-scours in the stratigraphy of Aeolis Mons, Gale crater. Fluvial Aeolian Interactions on Planetary Surfaces Workshop, The European Space Research and Technology Centre, Noordwijk, Netherlands.
- **A. L. Roberts**, S. Gupta, W. E. Dietrich, L. A. Edgar, W. Rapin, S. G. Banham, A. B. Bryk, J. M. Davis, E. Kite, A. Cowart, T. Kubacki, N. Moore, P. J. Gasda, J. R. Johnson, G. Caravaca. (2023). What depositional processes and palaeoenvironments formed the layered sulfate unit in Gale crater, Mars?: Insights from Marker Band valley. Lunar and Planetary Science Conference, The Woodlands, TX.
- **A. L. Roberts**, S. Gupta, P. Fawdon, S. Banham, J. Davis. (2022). Reconstructing the palaeoenvironment of the sediment fan at Oxia Planum, Mars. British Planetary Science Conference.

Chapter 2: Palaeo-scours within the layered sulfate-bearing unit at Gale crater, Mars: Evidence for intense wind erosion

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2.1 Introduction

Large-scale erosional scour structures preserved in the stratigraphic record are important signatures of landscape evolution on ancient planetary surfaces. Such surfaces can be formed by erosion and entrainment of sediment by both aqueous (Gibling, 2006; Priddy and Clarke, 2020) and aeolian processes (e.g. Blakey, 1988; Rubin and Hunter, 1982). Aeolian erosional structures may arise by autogenic processes such as scour pit migration during dune advancement (e.g. Cardenas et al., 2019; Kocurek and Day, 2018; Rubin and Hunter, 1982) or by allogenic processes such as base-level fall, changes in magnitude of sediment-transporting winds, changing sediment availability, and the position of the water table (Havholm and Kocurek, 1994; Kocurek and Day, 2018; Kocurek et al., 2001). Analysis of the geometry and distribution of ancient scour bedforms permits insight into past environmental evolution.

In this chapter, I investigate a series of decametres-wide scour-and-fill structures preserved in the Chenapau member of the Mirador formation which forms part of the stratigraphy of Aeolis Mons (informally Mount Sharp), Gale crater, Mars. Aeolis Mons is a ~5.5 km tall mound of layered sedimentary rocks that is interpreted to be a remnant of extensive crater-filling strata that formerly extended across the entire crater (e.g. Anderson and Bell III, 2010; Grant et al., 2014; Grotzinger et al., 2015; Thomson et al., 2011). Since 2012, the *Curiosity* rover has been characterising changing surface paleoenvironments in Gale crater through interrogation of this stratigraphic succession. The scour-and-fill structures that are the focus of this work occur within the Mirador formation and within a section interpreted to have formed in an aeolian sandsheet environment (Edgar et al., 2024; Roberts et al., 2023). Scour-and-fill structures can form through either aqueous or aeolian processes, and thus may

represent processes operating within the prevailing aeolian conditions, or brief returns to fluviolacustrine depositional conditions present within the lowermost strata observed along *Curiosity's* traverse (Grotzinger et al., 2015; Grotzinger et al., 2014; Williams et al., 2013). The nature of the erosional and depositional processes that created and infilled scours can provide insight into the climate of the region at the time of their formation such as wind conditions and the availability of surface water and sediment. Consequently, robust characterization and interpretation of the origin of these scours is required to understand the evolution of the Gale crater paleoenvironment ~3.6 billion years ago, a period that has been inferred to represent an important global climate change (e.g. Bibring et al., 2006; Carr and Head, 2010; Ramirez and Craddock, 2018). To address this, I conducted an analysis of the morphology and sedimentary character of these scour-and-fill structures to understand their formative processes and paleoenvironments.

The main objectives of this chapter are to: (1) describe the geometries of these decametre-wide erosional scours and their sedimentary infill, (2) describe the sedimentary character of the scour-and-fill structures and the host strata, (3) evaluate the mechanisms and possible controls on scour formation, and (4) discuss the implications for landscape evolution at Gale crater.

2.1 Geologic and Stratigraphic Context

Gale crater (Figure 2.1a) is a ~155 km-wide impact crater. It is located 5° south of the Martian equator, near the dichotomy boundary between the heavily cratered southern highlands and the relatively smooth northern lowlands (Tanaka et al., 2014). The crater has a model surface age of approximately 3.7-3.8 billion years (Ga) from crater counting of the impact ejecta blanket (e.g. Le Deit et al., 2013; Thomson et al., 2011) and subsequently infilled by sedimentary deposits over a 200 to 500 million year (Myr) period after crater formation (e.g. Grant et al., 2014; Thomson et al., 2011), including many periods of aeolian erosion (Dromart et al., 2021). The infilling material in Gale crater was later exhumed by predominantly aeolian processes (Anderson and Day, 2017; Day and Kocurek, 2016; Grotzinger et al., 2015; Kite et al., 2013b; Malin and Edgett, 2000) leaving a ~5.5 km tall central mound - Aeolis Mons (informally Mount Sharp) - exposing layered stratigraphy (Malin and Edgett, 2000; Anderson and Bell III, 2010) (Figure 2.1a). Orbital spectral observations of the north face of Aeolis Mons identified a vertical succession from clay-bearing strata transitioning to layered sulfate-bearing strata, which are unconformably overlain by anhydrous mineral-bearing strata (Figure 1b; Fraeman et al., 2016; Milliken et al., 2014). This clay-to-sulfate transition is found to be a variable and inconsistent transition, with discontinuous inclusions of clay-bearing strata amongst the layered sulfate-bearing strata from further orbital studies (Sheppard et al., 2021b).

As of middle 2024, *Curiosity* has traversed over 800 vertical metres of stratigraphy since landing at Gale in August of 2012 (Figure 2.2), documenting three major stratigraphic groups at the time of writing: the Bradbury, Mount Sharp, and Siccra Point groups. The Bradbury group consists of fluvial sandstones and conglomerates (Williams et al., 2013) which outcrop on the plains between the crater

rim and the base of the mound (Aeolis Palus) and are inferred to interfinger with strata of the Mount Sharp group, which form the sedimentary rocks in lower Aeolis Mons (Grotzinger et al., 2015). The Mount Sharp group consists of stratigraphy which bears hydrated minerals identified from orbit prior to landing (e.g. Milliken et al., 2010). These groups are unconformably overlain by the Siccar Point group. The Stimson formation is the only unit of the Siccar Point group identified so far (Banham et al., 2022; Watkins et al., 2022), and is composed of large-scale aeolian cross-bedded units (Banham et al., 2018) although minor soft sediment deformation structures show that near-surface fluid was present before lithification (Banham et al., 2024a; Banham et al., 2024b).

The Mount Sharp group, currently documented as approximately 700 metres thick at the time of writing, is subdivided into three formations (Figure 2.2). Exposed at the base of the mound is the Murray formation which comprises clay-bearing strata with lacustrine mudstones and subordinate aeolian and fluvial sandstones (Edgar et al., 2020; Grotzinger et al., 2014; Rivera-Hernández et al., 2019; Stack et al., 2019). It is conformably overlain by the Carolyn Shoemaker formation which contains interbedded mudstones and cross-bedded sandstones interpreted as an intermittent fluvial and lacustrine environment (Caravaca et al., 2022b; Cardenas et al., 2022; Fedo et al., 2022; Gwizd et al., 2024). Overlying the Carolyn Shoemaker formation, the Mirador formation is comprised of the layered sulfate-bearing strata seen from orbit (Fraeman et al., 2016).

The Mirador formation is subdivided into a series of members (Figure 2.2). The members which comprise the base of the Mirador formation contain large sets of cross-stratification which are attributed to the migration of aeolian dunes (Rapin et al., 2021). These metre-scale cross-sets occur throughout the Dunnideer, Port Logan, and Contigo members of the Mirador formation with varying degrees of diagenetic alteration (Edgar et al., 2024; Rapin et al., 2023b). Within the Contigo member, there is evidence for intermittent lacustrine and fluvial deposition through metre-scale resistant lenses (Caravaca et al., 2022a; Gupta et al., 2023b). The Contigo member is overlain by the Catrimani member which exhibits planar, laterally extensive laminae interpreted to be from the migration of wind ripples in aeolian sand sheets (Edgar et al., 2024; Roberts et al., 2023). Sharply overlying the Catrimani member is a thin (~0.5 m thick) member called the Amapari member (e.g. Mondro et al., 2025a; Mondro et al., 2025b; Thompson et al., 2024). This member is dark-toned, distinctive and can be seen from orbit as a marker band used in geomorphic mapping attempts (Anderson and Bell III, 2010; Milliken et al., 2010; Thompson et al., 2024; Thomson et al., 2011; Weitz et al., 2022). This member contains symmetrical ripples interpreted to be formed in a shallow lake (Gupta et al., 2023a; Mondro et al., 2025a; Mondro et al., 2025b; Weitz et al., 2023). The Amapari member is overlain by the Chenapau member which consists of planar-parallel strata which are interstratified with the large-scale scour-and-fill structures reported here.

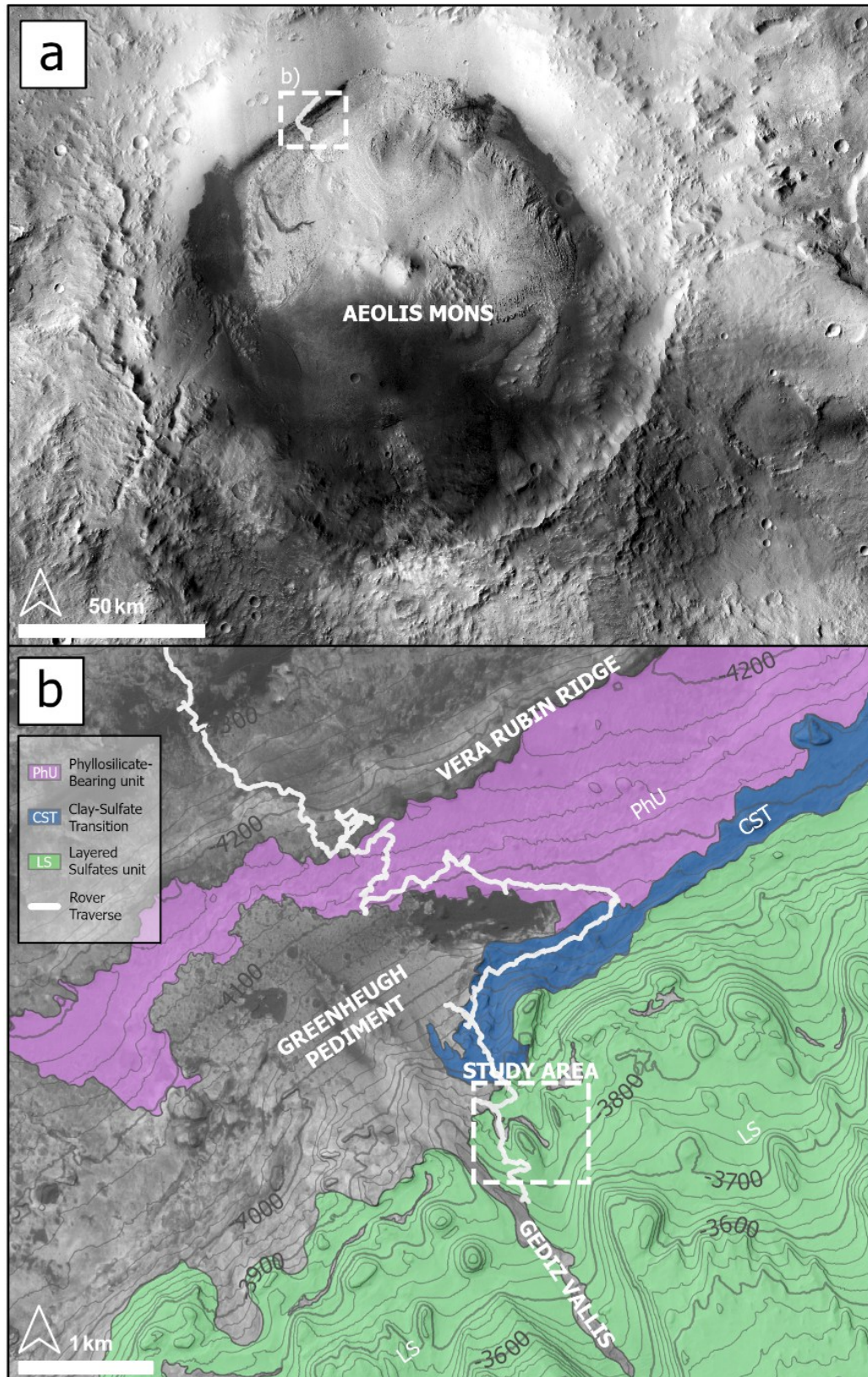


Figure 2.1: Study location at Gale crater. (a) Mars Reconnaissance Orbiter (MRO) Context Camera (CTX) mosaic (~5 m per pixel resolution; Dickson et al., 2018) showing Gale crater and Aeolis Mons. White dashed box shows Curiosity's exploration extent. (b) CTX mosaic overlain by a digital terrain model (DTM) hillshade and ~20 m contour intervals derived from Mars Orbital Laser Altimeter (MOLA) and High Resolution Imaging Science Experiment (HiRISE) data (~1 m per pixel resolution; Calef and Parker, 2016; Smith et al., 2001), shows the regional extent Curiosity's traverse within Gale crater until Sol 3749. Major Mount Sharp orbital units modified from Fraeman et al. (2016) on the northern flank of Aeolis Mons are displayed for context. White dashed box shows the study area at the base of the layered sulfate unit within Marker Band valley (Sols 3576 to 3749). Image credit: NASA/JPL/MSSS/The Murray Lab.

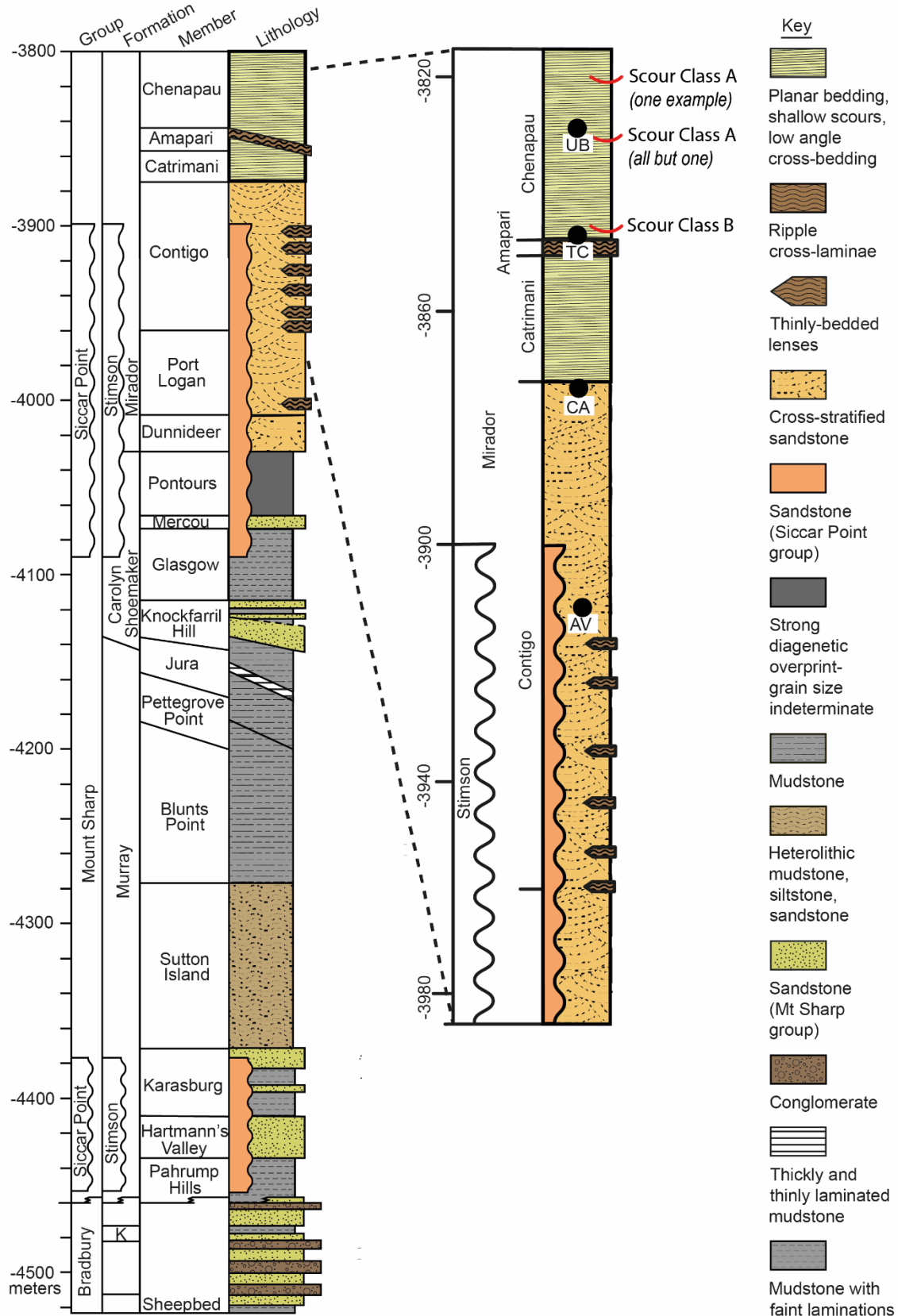


Figure 2.2: Stratigraphic column modified from the Mars Science Laboratory (MSL) Team's Sedimentology and Stratigraphy Working Group displays the vertical extents of the Bradbury, Mount Sharp and Siccar Point groups as encountered along the rover traverse. The inset stratigraphic column illustrates the vertical extents of the Contigo, Catrimani, Amapari, and Chenapau members of the Mirador formation, providing the stratigraphic context for the scour-and-fill structures in Marker Band valley. These structures are highlighted in red. Drill holes AV – Avanavero, CA – Canaima, TC – Tapo Caparo, and UB – Ubajara are also shown for reference.

2.3 Data and Methods

2.3.1 Instruments

The architecture of the scours primarily was investigated using the Mastcam, Navcam and ChemCam instruments aboard the *Curiosity* rover.

The two Mast Cameras (MastCams - left and right; Bell et al., 2017; Malin et al., 2017), are located at the top of the rover's remote sensing mast (RSM) - ~ 1.9 m above ground level - and provide high-resolution colour and stereo photomosaics. The cameras are separated by ~ 24.2 cm which allows for stereo data collection at the spatial sampling of the M34 camera, which is 0.22 mrad/pixel (~ 0.45 mm/pixel at 2 m viewing distance). The 0.074 mrad/pixel sampling of the M100 camera provides ~ 0.15 mm/pixel at 2 m viewing distance. Mastcam data products were processed (Table II.2, Appendix II) and mosaicked to optimise both visibility and accuracy of the sedimentary architecture for analysis. Mastcam spatial sampling of scour image coverage ranged from 17 to 5.8 mm/pixel depending on the image range (Table II.5, Appendix II).

The Navigation cameras (Navcams; Maki, 2018; Maki et al., 2012) are also located near the top of the RSM and provide 360°, monochromatic (greyscale) context stereo mosaics at every rover waypoint providing additional spatial context for Mastcam mosaics. The left and right navigation cameras both have a fixed focal length of 14.67 mm and provide 0.82 mrad/pixel. With an interocular length of ~ 42.4 cm, stereo three-dimensional mesh products were derived (Alexander et al., 2006) typically accurate to a distance of ~ 30 -50 m from the rover and provided three-dimensional data for scours within range.

ChemCam (Wiens et al., 2012) has the capability to take long distance (LD), high-resolution, monochromatic image mosaics with the Remote Micro-Imager (LD RMIs; Le Mouélic et al., 2015; Le Mouélic et al., 2019). With a narrow field of view at a fixed focal length of ~ 700 mm and 0.0205 mrad/pixel, LD RMIs were sometimes used to supplement Mastcam data where instrument pointing safety restrictions allowed.

The Mars Hand Lens Imager (MAHLI) is attached to the turret on *Curiosity*'s arm, and can provide detailed colour images with spatial sampling up to ~ 14 μ m per pixel (Edgett et al., 2012). Typical MAHLI imaging procedures involve a suite of images captured at nominal toolframe distances of 1 cm (spatial sampling of 16 to 21 μ m per pixel), 5 cm (spatial sampling 31 μ m per pixel), and 25 cm (spatial sampling of 100 μ m per pixel) (Yingst et al., 2016). A second observation suite, referred to as a "dog's eye," places MAHLI orthogonal to the sedimentary structure within the outcrop. Therefore, MAHLI can be used to resolve fine-scale sedimentary structures and textures at a hand-lens scale with the ability to typically resolve grain sizes of coarse silt, ~ 62.5 μ m, or larger (Edgett et al., 2015; Yingst et al., 2016).

The rover-derived data products were supplemented with ~ 25 cm/pixel orbital images from the High Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007). For three-dimensional

data, a ~1 m/pixel Digital Terrain Model (DTM) generated from HiRISE stereo pairs and registered to the Mars Orbiter Laser Altimeter data products (Calef and Parker, 2016; Smith et al., 2001) was used (see Appendix III.1). These data products aided in photogeological mapping, rover and outcrop localization, and strategic planning of rover data collection sites (Figure I.1, Appendix I).

2.3.2 Data Collection and Imaging Strategy

Between Sols 3576 and 3749 (the number of Martian days since *Curiosity* landed in Gale crater), the rover traversed across Marker Band valley. This valley is topographically defined by the buttes Orinoco, Chenapau, and Bela Vista, as well as a ledge forming the western valley flank, all of which host nine of the scours described in this chapter (Figures 2.3 and 2.4). Additionally, another scour was identified during a rover detour up to Gediz Vallis Ridge from this valley. When time, data volume, and energy constraints allowed (e.g. Vasavada, 2022), mosaics were planned around the rover's closest approach along the Mount Sharp Ascent Route (MSAR) and when viewsheds were orthogonal to the feature of interest to avoid distortion from image foreshortening effects. Seven key M34 and M100 stereo mosaics of the scour-and-fill structures were taken (Figure I.1, Appendix I; Table II.2, Appendix II). Due to lighting and instrument safety constraints, only two LD RMIs were taken of these scours (Table II.3, Appendix II). All outcrops, geologic units, and rock targets investigated by *Curiosity* are informally named by the MSL team based on naming conventions detailed in Vasavada et al. (2014).

Although the rover traverse did not intersect any scour-and-fill structures, the traverse crossed laterally equivalent strata in two separate traverse sections - along the Canta detour and again along the main traverse route west of Chenapau (Figure 2.3) - which provided nearfield (<10 m distance) observations of the strata to enable reconstruction of depositional processes. The scour-hosting facies were characterized by a transition from relatively “dark-toned” and “light-toned” strata (Figure 2.4d). These stratigraphic transitions were correlated and mapped around the two scour-hosting buttes and onto the projected rover traverse (Roberts et al., 2023). This map of the stratigraphic contact (Figure I.2, Appendix I) was then used to visit the lateral equivalent of the scour-hosting facies with the rover for on-ground analysis.

2.3.3 Data analysis

As the range to various sections of the scours varied relative to the rover (Figure I.1, Appendix I; Table II.5, Appendix II), different approaches were required to quantitatively assess the scours. The three-dimensional Navcam mesh (Alexander et al., 2006) was used to provide width and thickness data of the scours at distances less than ~50 m. At distances further than 50 m, the scours were mapped using *ArcGIS Pro 3.1.0* onto the HiRISE orthomosaic and used as a base to correlate features of interest visible in both the HiRISE and Mastcam mosaics (Table II.5, Appendix II). The GIS projected viewsheds using the HiRISE DTM based on the rover co-ordinates and imaging specifications could also be used to help correlate features of interest to the Mastcam mosaics (see Appendix III.2). The

mapped scours could then be quantitatively assessed using the HiRISE DTM. To account for foreshortening when measuring apparent dips, I only measured the parts of the image mosaic oblique to the scours (Figure I.1, Appendix I).

Analysis of the host stratigraphy was predominantly performed using stereo coverage derived from M34 and M100 data products and more detailed coverage from M100 (a list of Mastcam mosaics used in this chapter is listed in Table II.2, Appendix II). MAHLI images help resolve fine-scale sedimentary structures in the host stratigraphy for a more detailed facies characterization (a list of MAHLI images used in this chapter is listed in Table II.4, Appendix II). Measurements of laminae thickness could be measured based on visual inspection of MAHLI images orthogonal to bedding as described by Edgar et al. (2020), using the open-source image processing program, ImageJ (Schneider et al., 2012).

2.4 Stratigraphic context of scour-and-fill structures

Within the Chenapau member (Figure 2.2), ten decametre-wide scour surfaces were identified truncating low-angle, planar strata. These scours occur associated with three visually defined stratigraphic units within the Chenapau member. Most scours are incised into a dark-toned unit and are overlain by a light-toned unit (Figure 2.3; 4). The remaining scours are embedded within a light-toned unit where diagenetic overprinting has obscured the sedimentary textures of the host stratigraphy. Hereafter, this will be referred to as the “diagenetically overprinted unit.” Here, I provide a brief description of these units from both the outcrop scale (50-300 m range) and rover nearfield scale (< 10 m range) and present our interpretation of their depositional setting.

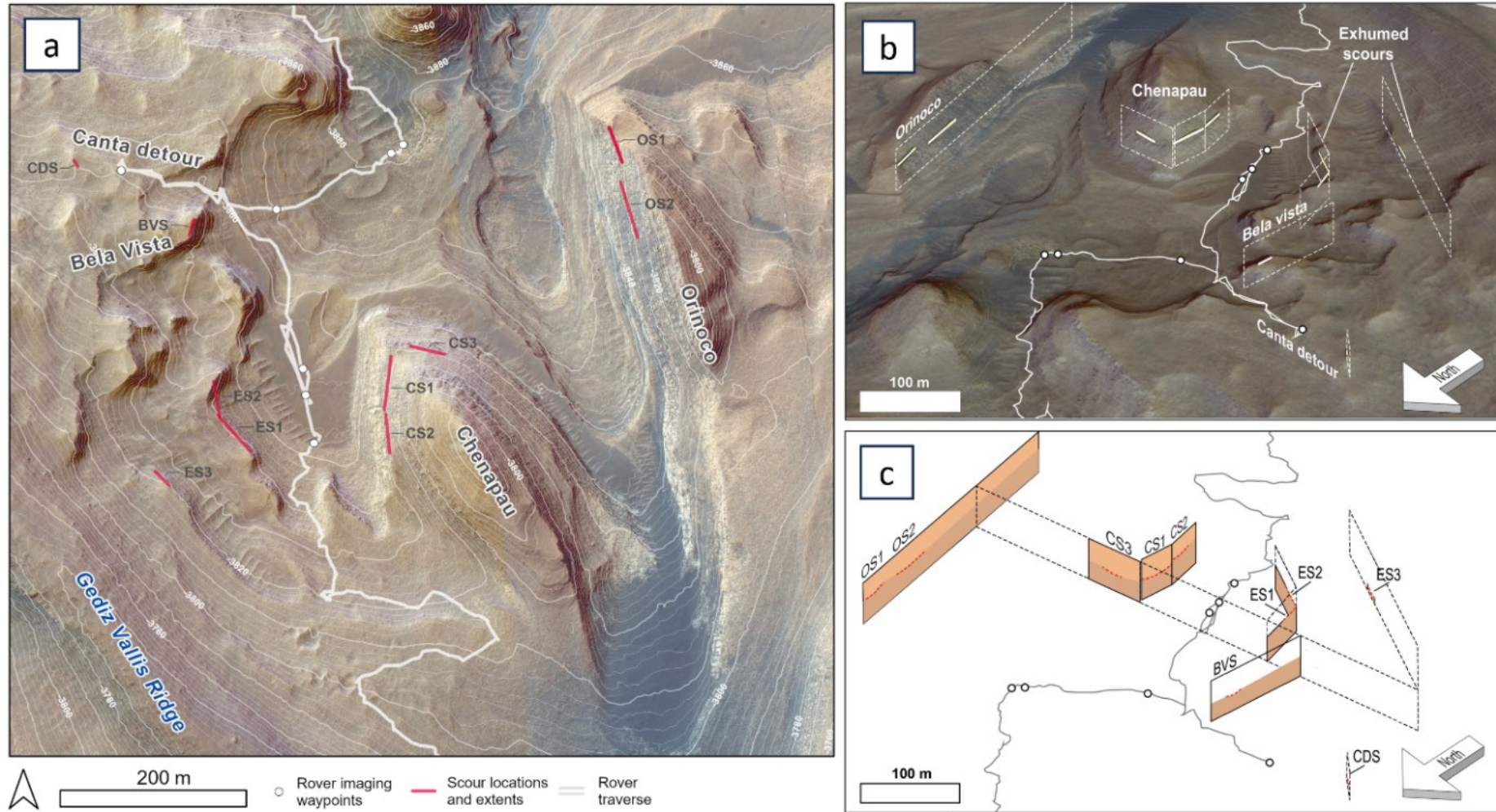


Figure 2.3: Context of scour-and-fill structures. (a) High Resolution Imaging Science Experiment (HiRISE) mosaic (~25 cm per pixel; Calef and Parker, 2016) overlain by a MOLA and HiRISE derived digital elevation model hillshade shows the Marker Band valley region of the rover traverse (from Sols 3576 to 3749). Red lines show scour locations, orientations, and extents with shorthand labels (longhand names can be found in Table 2.1). White line shows Curiosity's traverse. Black-rimmed circles show the key imaging locations. (b) ArcGIS Pro HiRISE scene-view of the location, extent, and orientations (see Table 2.1 for numeric orientations) of the scour-and-fill structure mosaics included in this chapter (dashed white lines). Thick white lines delineate the scours, gray line delineates the rover's traverse, black-rimmed white circles show key imaging locations, and the dark-/light-toned shading shows the dark- and light-toned unit contact boundary. Image credits: NASA/JPL/University of Arizona

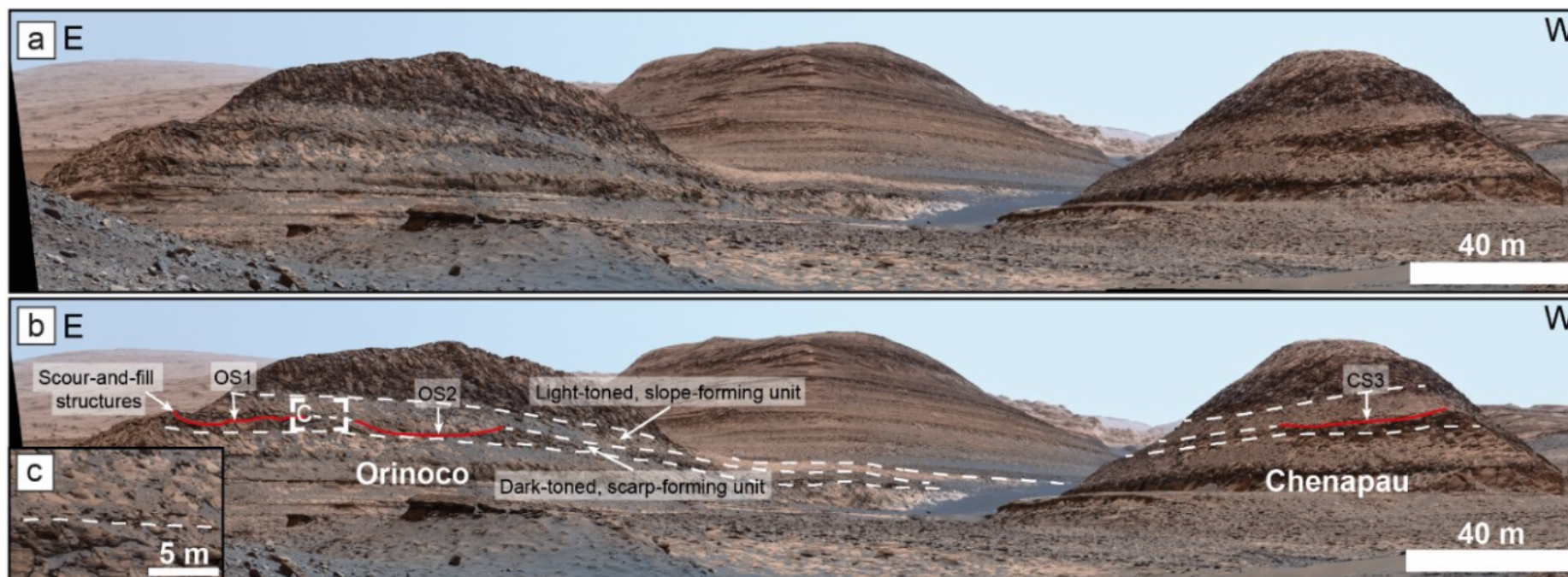


Figure 2.4: (a) Overview mosaic showing Orinoco and Chenapau buttes in the Marker Band valley study area (Mastcam mosaic MR_mcam102554, acquired on Sol 3562, at Sol 3650 end-of-drive location). (b) Interpreted image showing the scour-enclosing unit boundaries (dashed white lines) and the scour-and-fill structures (red lines). (c) Detail showing the stratigraphic context of the dark- and light-toned units. Image credits: NASA/JPL-Caltech/MSSS

2.4.1 Observations

Dark-toned stratigraphic unit

A series of scours incise a ~6 m thick interval of visually dark-toned, scarp-forming strata at a single stratigraphic level between elevations -3,833 and -3,828 m. At an outcrop scale, this interval is found to be comprised of cm-scale sets of recessive-resistant packages of planar strata (Figure 2.4c). The dark-toned stratigraphic interval was encountered along the rover's traverse from Sols 3801 to 3817, between elevations -3832 and -3826 m (Figure I.2, Appendix I). Nearfield imaging shows 1-8 cm thick recessive planar packages comprised of planar, laterally extensive laminae that are <1 mm thick and which can be traced across blocks for over ~3 m. Planar, weathering-resistant laminae separate these recessive sets and protrude out from bedrock targets (Figure 2.5a). Isolated sets of indurated laminae ~1 cm thick are observed within this unit.

Light-toned stratigraphic unit

Overlying both the dark-toned stratigraphic interval and the scours is a light-toned interval measuring ~8 m thick. Along section, this interval appears predominantly structureless although some sections show evidence of planar stratification (Figure 2.4). The light-toned stratigraphic interval was encountered along the rover's traverse from Sols 3812 to 3867, between elevations -3826 and -3814 m (Figure I.2, Appendix I). Nearfield observations show the interval is composed of planar and laterally extensive laminae (Figure 2.5b, and d). Individual laminae - traced laterally for ~3 m across outcrops - demonstrate planar geometry with no thickness variation. Each individual lamina is composed of a light-toned/resistant and dark-toned/less resistant component resembling a pin-stripe pattern (Figure 2.5d). Measured sections of this stratigraphic interval in MAHLI images show remarkably even-in-thickness laminae, with an apparent thickness of ~0.76 mm and a standard deviation of 0.058 mm (Figure I.3, Appendix I). No other sedimentary structures were identified from rover observations.

Diagenetically overprinted unit

Along the "Canta detour" (Figure 2.3), the original sedimentary structures and textures are difficult to resolve. The rocks are heavily weathered and often have a nodular surface texture with cross-cutting veins (Figure 2.5C) which I interpret to be from diagenetic overprinting.

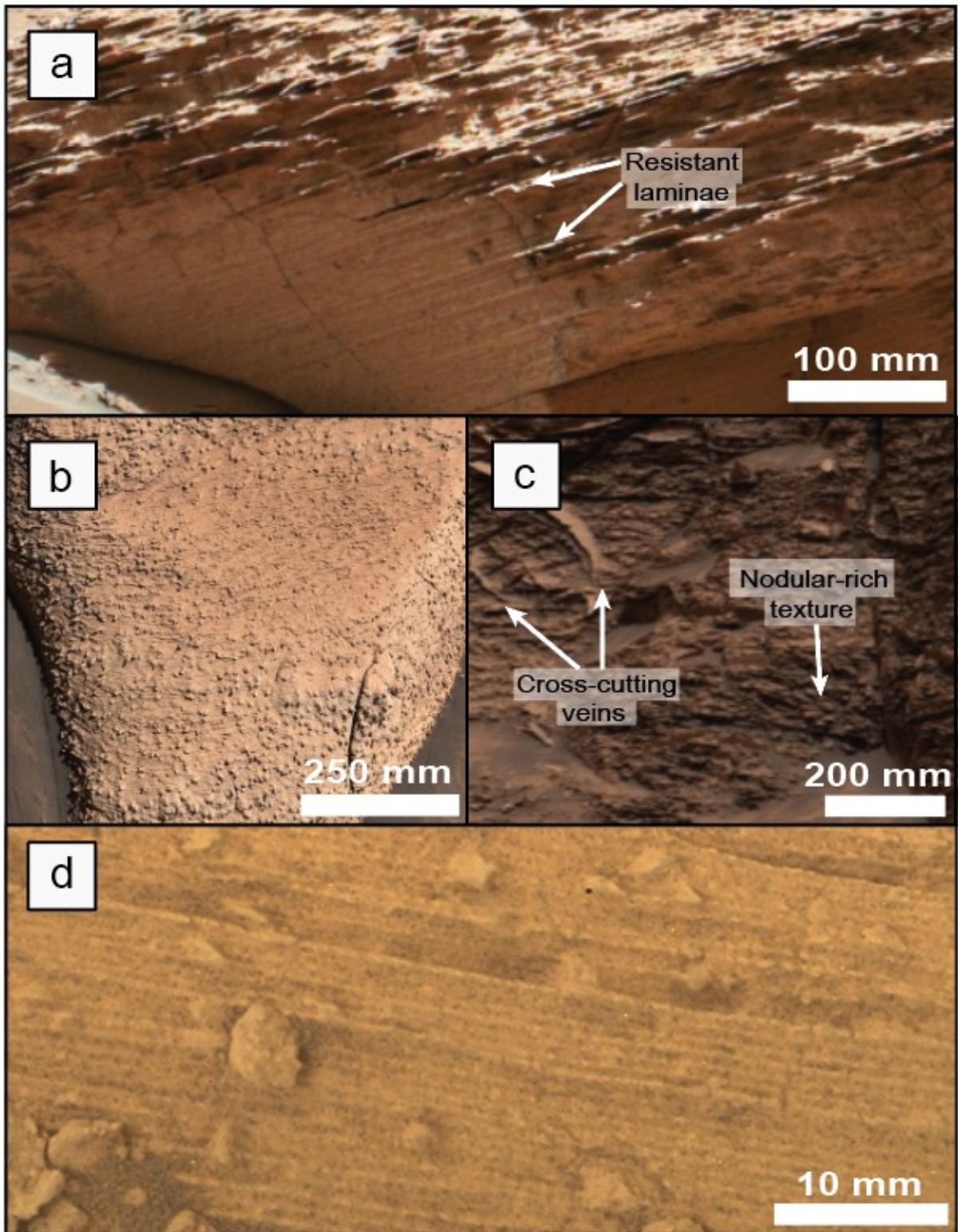


Figure 2.5: Examples of selected facies characteristic of the scour-enclosing strata. (a) Mastcam photomosaic of the dark-toned unit with resistant laminae separating cm-scale thick sets of light-toned, recessive laminae (Mastcam mosaic MR_mcaml03951, acquired on Sol 3807, at Sol 3805 end-of-drive location). (b) Mastcam mosaic of the light-toned unit (Mastcam mosaic MR_mcaml04131, acquired on Sol 3845, at Sol 3843 end-of-drive location). (c) Mastcam mosaic of the diagenetically overprinted unit with a nodular-rich texture and cross-cutting veins (Mastcam mosaic MR_mcaml03061, acquired on Sol 3648, at Sol 3646 end-of-drive location). (d) Part of a MAHLI dogs eye mosaic of the light-toned unit showing the laterally extensive, light-toned, planar laminae (Kastria Spring target, 25 cm stand off). Image credits: NASA/JPL-Caltech/MSSS.

2.4.2 Interpretation

The uniformity, lateral persistence, and planar nature of the laminae comprising the light-toned strata resemble laminae formed by the migration of wind ripples on Earth (e.g. Ferronatto et al., 2019; Kocurek and Nielson, 1986) and Mars (Banham et al., 2018). On Earth, wind ripples are often identified by inverse grading, where fines and silts are sequestered in ripple troughs and protected from migration of the coarser grained ripple-component (Fryberger and Schenk, 1988; Mountney, 2006a). Inverse grading is often difficult to identify on Mars as the finer particles are smaller than image resolution limits (e.g. Banham et al., 2022; Stack et al., 2019). However, a thin layer of fine particles often has a distinct tone (Mountney, 2006a) or exhibits preferential cementation (Fryberger and Schenk, 1988) compared to the rest of the lamina, creating a distinct pin-stripe pattern characteristic of aeolian wind-ripples (Fryberger and Schenk, 1988; Mountney, 2006a). Our observations of this pattern within the light-toned stratigraphic interval suggest the strata originated from wind-ripple migration (Figure 2.5). Furthermore, laminae formed by wind-ripple migration can be evenly laminated and have similar thicknesses (e.g. Banham et al., 2018; Fryberger et al., 1979; Fryberger and Schenk, 1988; Hunter, 1977), like those identified in the light-toned unit (Figure 2.5b, and d). The low-angle, flat depositional habit of the laminae and their lateral continuity over ~2 m leads me to interpret these deposits as products of wind-ripple migration in an aeolian sand-sheet environment: a flat, arid environment covered in wind-blown sand (e.g. Chakraborty and Chakraborty, 2001; Ferronatto et al., 2019; Grotzinger et al., 2005; Kocurek and Nielson, 1986). The absence of dune cross-stratification may indicate a process inhibiting dune development such as the presence of near-surface water (e.g. Kocurek and Nielson, 1986), or a small degree of cementation by salt (Nickling, 1984).

I interpret that the apparently dark-toned appearing strata were similarly deposited in an aeolian environment due to the recessive sets of planar, even-in-thickness, and laterally extensive laminae. The prominent laminae are interpreted to be more indurated than the light-toned recessive sets, providing additional weathering resistance. This induration is attributed to processes post-dating scour formation such as early cementation due to varying fluid retention properties, the presence of fine-grained clays trapped in the inverse graded wind-ripple laminae (Cain and Mountney, 2009), or post-depositional diagenesis due to variations in permeability influencing fluid flow. In aeolian rocks, these properties are thought to be influenced by changes in grain size, grain packing, wrought by variations in sediment transport process (e.g. Chandler et al., 1989; Fryberger and Schenk, 1988; Prosser and Maskall, 1993; Taggart et al., 2010). The presence of resistant laminae is interpreted as the cause of the darker tone of this unit at outcrop scale due to centimetre-scale self-shadowing. Both the dark-toned and light-toned strata are interpreted as aeolian facies suggesting an overall aeolian setting within which the scours formed.

2.5 Description of Scours-and-fill Structures

Here, I describe ten scour-and-fill structures identified within the Chenapau member of the Mirador formation. I broadly divide these structures into two classes, A and B, based on both morphology and infill characteristics. Class A consists of nine large scour-and-fill structures with a simple, symmetric morphology and light-toned draping infill. Class B consists of one smaller scour-and-fill structure with a dark-toned infill and winged-shape feature. Within class A, the nine scour-and-fill structures exhibit slight variations in their morphology and infill architecture. Five scour-and-fill structures have a simple, symmetric morphology and light-toned infill; two scour-and-fill structures have a complex scour and infill architecture; and three scour-and-fill structures are heavily exhumed, preserving little to no infill. Four structures preserve their complete geometry and infill architecture and the remaining six structures are partially preserved (Table 2.1). The locations of these scours are shown in Figure 2.3. The geometry and infill stratigraphy of each of these scour-and-fill structures are described below.

Class	Scour-and-fill structure	Apparent width (m)	Apparent thickness (m)	Elevation (m)	Apparent dip of scour margin relative to underlying strata	Cross-section azimuth
A	Chenapau Scour One (CS1)	~61	~4	~-3832 to -3836	~14-16°	8°
	Chenapau Scour Two (CS2)	~41	~3	~-3834 to -3829	~16-18°	355°
	Orinoco Scour Two (OS2)	~62	~4	~-3828 to ~-3836	~8-10°	343°
	Chenapau Scour Three (CS3)	~37	~2	~-3832 to ~-3834	<14°	103°
	Orinoco Scour One (OS1)	~40	~3	~-3832 to ~-3838	Section 2.2.1: <21°, Section 2.2:	340°
		~Part 1= ~21			<11°	
		~part 2= ~12				
		~part 3= ~4				
	Bela Vista Scour (BVS)	~16*	~<1*	~-3842 to -3843	<i>Unidentifiable</i>	199°
	Exhumed Scour One (ES1)	~40*	~1*	~-3838 to -3839	12-13°	140°
B	Exhumed Scour Two (ES2)	~49*	~3*	~-3838 to -3841	<16°	148°
	Exhumed Scour Three (ES3)	~19*	~4*	~-3819 to -3823	<i>Unidentifiable</i>	134°
	Canta Detour Scour (CDS)	~7	~<0.5	~-3850	<i>Unidentifiable</i>	150°

Table 2.1: Detailed measurements of the elevations, apparent scour-and-fill structure widths, apparent scour thicknesses, and apparent dips of scour margins for scours in classes A and B. *Partially preserved

2.5.1 Class A Scour-and-fill structures with a simple, symmetric morphology and light-toned infill

2.5.1.1 Observations

Chenapau Scour 1 (CS1)

CS1 is present on the north side of the western face of Chenapau butte and exhibits a cross-sectional profile measuring ~60 m wide and ~4 m deep (Figure 2.6). In profile, CS1 has concave-up and broadly symmetric margins which truncate the underlying planar strata (Figure 2.7c) with a subhorizontal planar central base section (Figure 2.7d). The scour only shallowly erodes into the underlying strata, with a maximum apparent margin dip of ~14° to ~16°. The scour occurs between -3836 and -3832 m elevation.

This scour morphology is consistent and simple; there are no deviations exceeding centimetre-scale amplitudes in the concave-up morphology of the scour paleo-surface. Scour fill consists of two components (Figure 2.6c): a basal section of low-angle inclined bedsets which conformably drape across the scour surface and an overlying section of light-toned strata lacking distinct bounding surfaces.

The basal section is distinguished by 1-2 m thick bedsets of light-toned strata separated by prominent bounding surfaces that form resistant bands that drape the scour's paleo-surface (Figure 2.7). The lowermost bedset is inferred to downlap onto the flat central portion of the basal scour contact, while the upper sets are inferred to onlap onto the scour's southern margin (Figure 2.6). The strata within each set are light-toned, broadly conforming to the lower bounding surface, and are truncated by overlying bounding surfaces (Figure 2.7e). The lowermost sets exhibit local undulations within the middle-basal portion of the scour (Figure 2.7d) which contrasts with the overlying and underlying bounding surfaces. In one location these undulating strata are truncated by the upper bounding surface (Figure 2.7e).

The light-toned fill section contains strata with low-angle dips which gradually decrease stratigraphically upwards until this section is indistinguishable from the light-toned unit (Section 2.4) in non-scoured sections of the stratigraphy (Figure 2.6b and c). The resistant bands are also gradually lost.

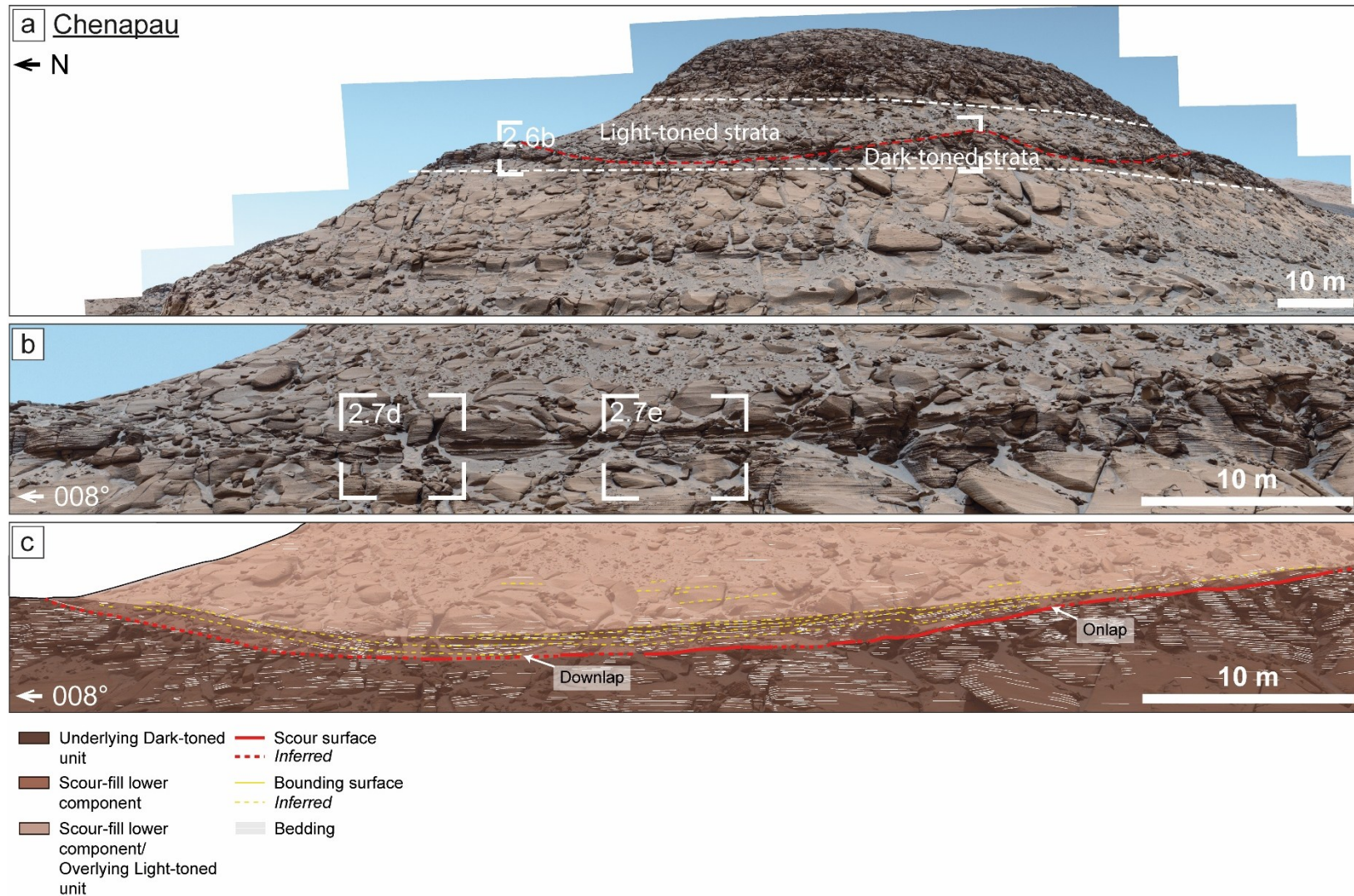


Figure 2.6: Chenapau butte mosaic showing Chenapau Scour 1 (CS1). (a) Mastcam mosaic of CS1 illustrating stratigraphic contacts (white dashed lines) and locations of erosive bases to scour-and-fill structures (red dashed lines) (Mastcam mosaic MR_mc103485 acquired on Sol 3727, at Sol 3724 end-of-drive location). (b) Detail of CS1 uninterpreted. (c) Interpretation of CS1 showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop. Image credits: NASA/JPL-Caltech/MSSS.

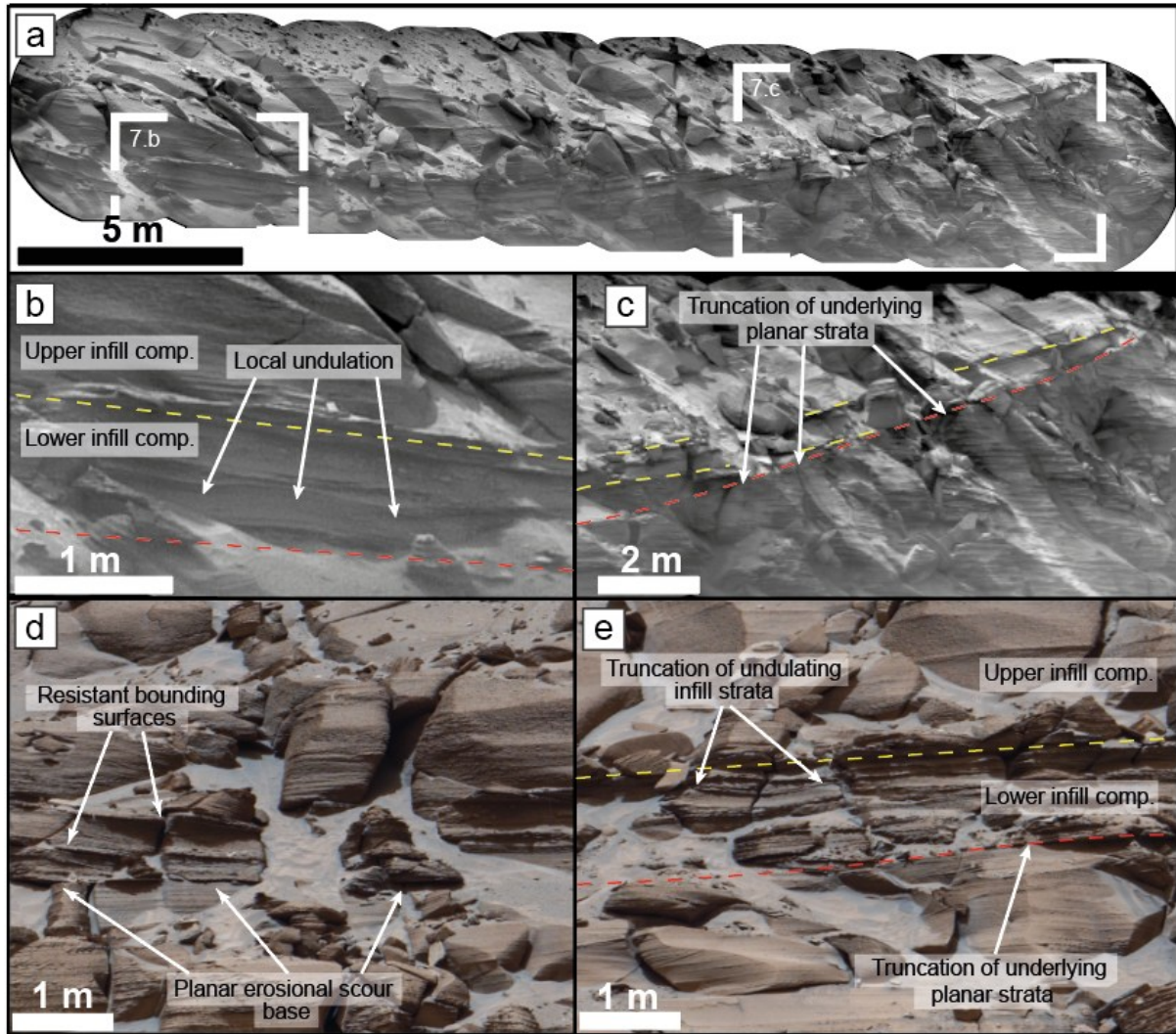


Figure 2.7: Details of Chenapau Scour 1 architecture. (a) A Long Distance Remote Micro-Imager (LD RMI) mosaic of Chenapau Scour 1 (ccam 05671, acquired on Sol 3671, at Sol 3667 end-of-drive location). (b) Detail showing the local undulation in scour infill bedset. (c) Detail showing the scour surface truncating the underlying bedding. (d) Mastcam detail of 6.b showing the scour surface is locally planar and parallel to the underlying bedding. (e) Mastcam detail of 6.b showing that the scour fill bedding undulates and is truncated by overlying scour fill bedding. Image credits: (a-c) NASA/JPL/LANL/CNES/IRAP ; (d-e) NASA/JPL-Caltech/MSSS.

Chenapau Scour 2 (CS2)

CS2 is present on the southern side of the western face of Chenapau and measures ~41 m in width and ~3 m in depth. The scour occurs between -3834 and -3829 elevation. It has a relatively simple, symmetric concave-up profile, with scour margins exhibiting maximum apparent dips of 16-18° grading onto a central planar base (Figure 2.8). The planar base locally deviates from a concave-up morphology with a centimetre scale erosional low in the scour surface (Figure 2.8d).

The scour infill is composed of two components. The lower fill section, ~0.5 m thick, is composed of bedsets of light-toned strata conforming to the northern margin and scour base. Bounding surfaces that separate these bedsets appear as prominent bands and become less prominent along the northern margin, although strata remain conformable to the base (Figure 2.8c). Bedsets within the central portion stack both vertically and laterally northwards along dip. Bounding surfaces truncate underlying bedsets in two locations (Figure 2.8d) and continue downdip until they intersect the basal contact and incise

planar strata underlying the scour. The strata within the bedsets are light-toned, recessive, and conform to the bounding surfaces. Undulation of the bedsets or internal strata are not observed in this scour-and-fill structure. The upper fill section of CS2 is recessive and light-toned and is distinguished from the lower fill section by the lack of distinct sets. This section becomes progressively more planar vertically upwards until it is indistinguishable from the overlying “light-toned” unit.

Orinoco Scour 2 (OS2)

OS2 occurs on the western face of Orinoco butte and exhibits a cross-sectional profile measuring ~62 m wide, ~4 m thick. It is a single, simple, concave-up scour surface with a symmetric and uniform morphology (Figure 2.9). The scour margins exhibit maximum apparent dips between 8 and 10 degrees relative to the underlying planar strata. The central base is parallel to the underlying strata. This was the most distant (~300 m range) scour-and-fill structure imaged by the rover in this study, with spatial sampling at ~22.2 mm/pixel.

Lower and upper infill components are observed within OS2. The lower fill is composed of bedsets of light-toned, recessive strata separated by resistant bands conforming to the scour surface. The resistant bands are progressively lost southward across the southern scour margin. The bedsets themselves appear to downlap onto the central scour base, and onlap the southern margin, giving the appearance of stacking laterally to the south. Towards the northern margin, the scour appears overfilled and strata become planar and can be traced outside the confines of the scour. The upper fill is composed of light-toned strata with similar erosional expression to strata confined within bedsets in the lower scour fill. At this resolution, the upper fill strata appear structureless and indeterminate from the overlying light-toned unit.

Chenapau Eastern Face Scour (CS3)

CS3 occurs on the eastern face of Chenapau. Interpretation of this outcrop is challenging due to coverage by float blocks, coarse imaging scale, and a strongly foreshortened perspective. Truncation of the underlying strata was not directly observed; the scour is inferred from light-toned strata with apparent dips differing from the underlying, planar dark-toned strata (Figure 2.10). The erosional surface was inferred from the extent of the changes in apparent dip and measured to be ~37 m wide and ~2 m thick. The scour occurs between elevations ~-3834 m and -3832 m.

A two-component infill succession is observed. The lower component is comprised of inclined sets of light-toned strata divided by dark-toned bands. The upper component is light-toned and indistinguishable from the overlying unit (Figure 2.10). At the northernmost part of the butte, the light-toned strata dip southwards. Towards the south, strata first flatten out before becoming increasingly inclined northwards, giving the appearance of a concave-up scour fill similar to CS1 and CS2. Figure 2.10 shows a distinct surface between these inclined strata and the underlying dark-toned appearing strata at the centre which appear parallel to the underlying strata.

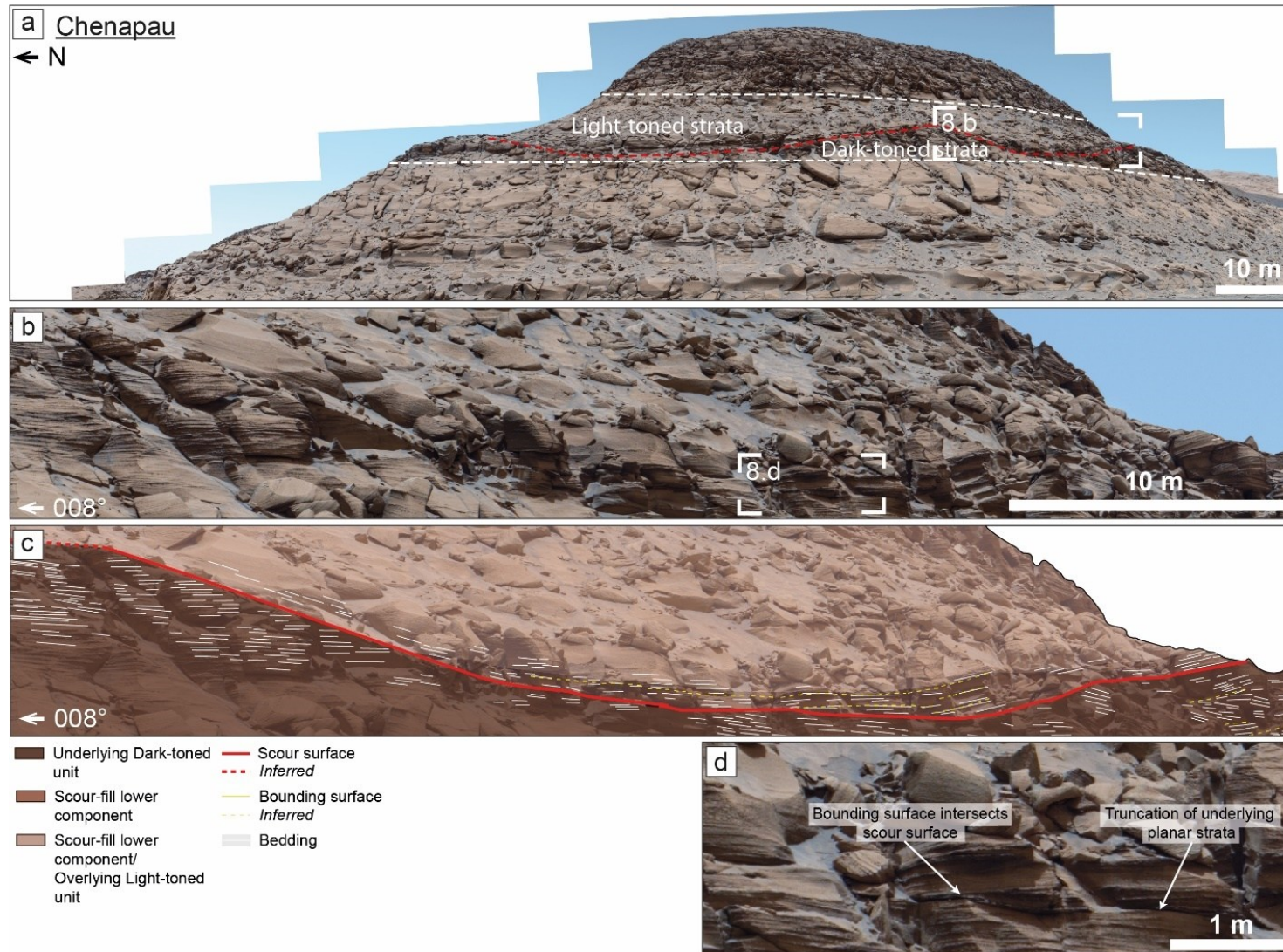


Figure 2.8: Architecture of Chenapau Scour 2 (CS2). (a) Mastcam mosaic of Chenapau butte illustrating the stratigraphic contacts (white dashed lines) and locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mcam103485 acquired on Sol 3727, at Sol 3724 end-of-drive location). (b) Detail showing CS2 unit uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop. (d) Detail showing intersection of the basal scour erosion surface by a bounding surface from within the scour fill. Image credits: NASA/JPL-Caltech/MSSS.

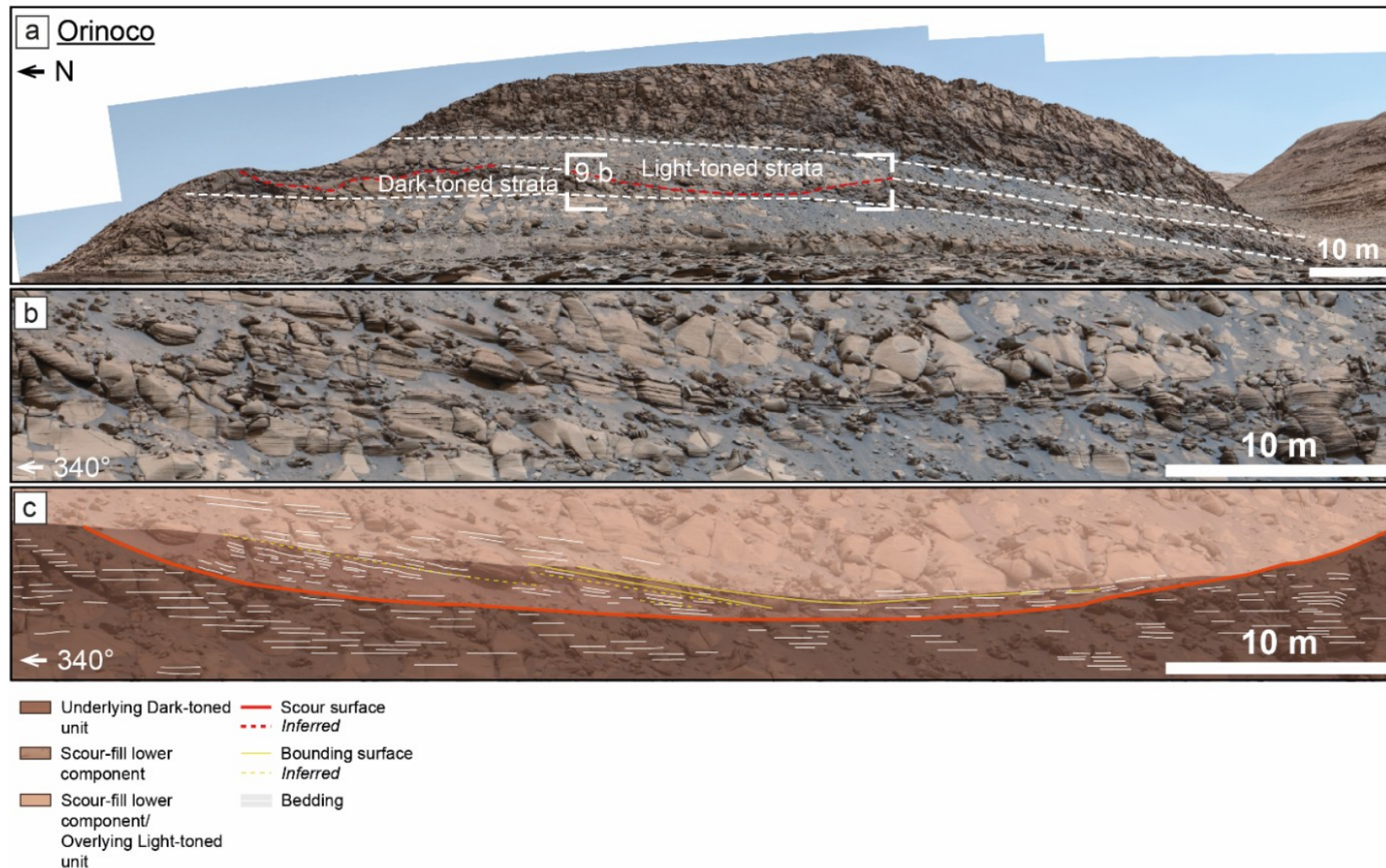


Figure 2.9: Architecture of Orinoco Scour 2 (OS2). (a) Mastcam panorama of the Orinoco butte illustrating the stratigraphic contacts (white dashed lines) and the locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mcam102672 acquired on Sol 3578, at Sol 3576 end-of-drive location). (b) Detail of OS2 uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop. Image credits: NASA/JPL-Caltech/MSSS.

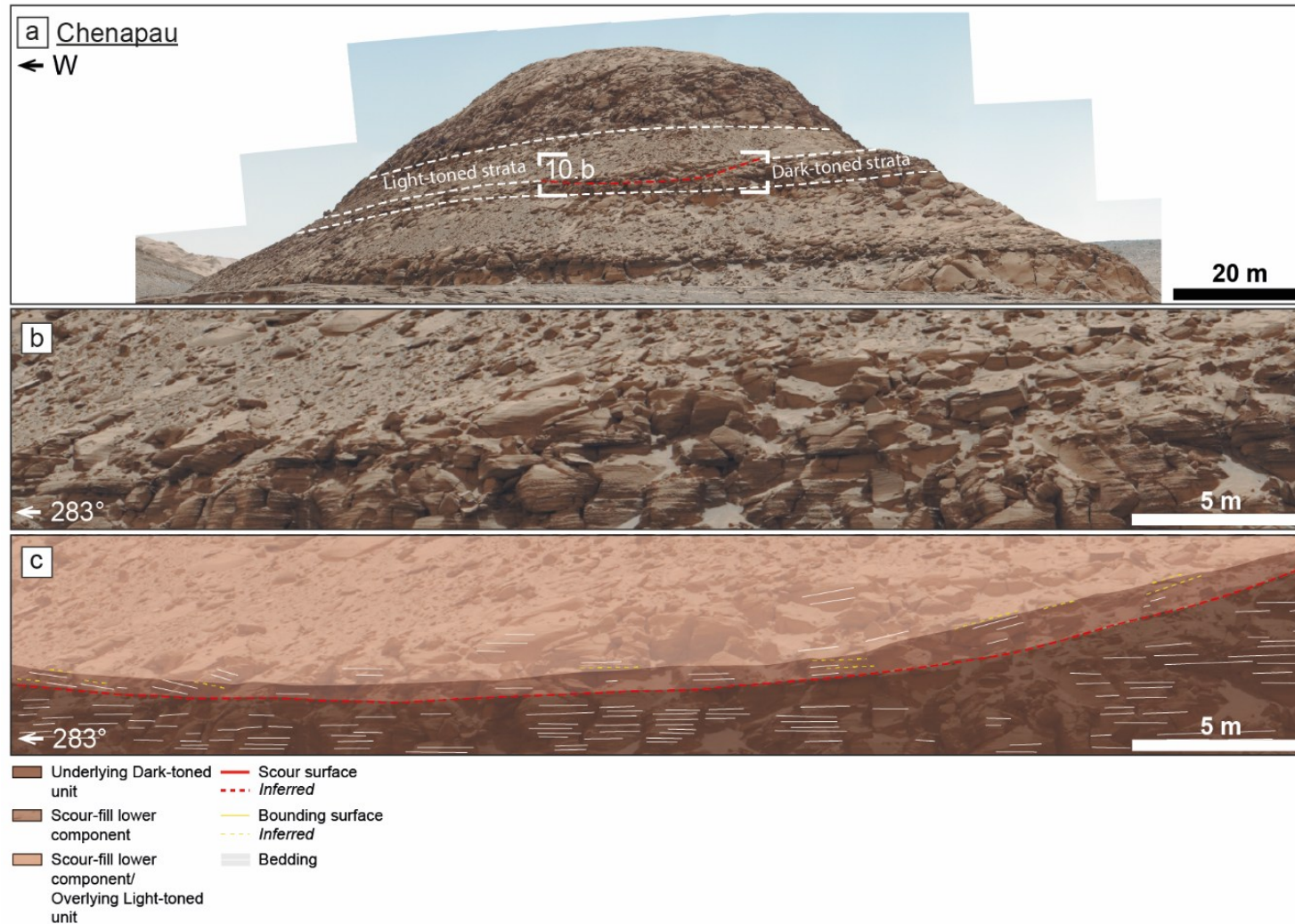


Figure 2.10: Architecture of Chenapau Scour 3 (CS3). (a) Mastcam panorama of the eastern face of the Chenapau butte illustrating the stratigraphic contacts (white dashed lines) and the locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mc102724, acquired on Sol 3589, at Sol 3579 end-of-drive location). (b) Detail of CS3 uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop. Image credits: NASA/JPL-Caltech/MSSS.

2.5.1.2 Interpretation

I find that CS1, CS2, CS3, and OS2 all share a uniform scour morphology with apparent width to thickness ratios (W/T) of > 15 and a similar two-component scour infill characteristic. The similarity of these scour-and-fill structures suggests shared erosional and depositional processes. As scour-and-fill structures CS1, CS2, CS3, and OS2 all occur within the same stratigraphic interval between the dark- and light toned stratigraphic units, within ~ 9 m of topographic elevation over a 0.25 km^2 area (Figure 2.4; Table 2.1), I interpret that they all formed during the same narrow temporal window.

The light-toned sets observed in the lower component of the scour fill closely resemble the upper component, with the key distinction being the presence of resistant bounding surfaces (Figure 2.7d). While local undulations in the lower component bedsets were observed, these only occur at the base of CS1 (Figure 2.7b, e). Thus, I interpret this undulation as a localized depositional texture, such as an expression of microtopography along the scour base, and not representative of typical class A infill. Truncation of these undulating bedsets from overlying bounding surfaces indicate that this process occurred contemporaneously with infill deposition. While the resistant bands are often found in association with bounding surfaces, they are also found to separate the sets of light-toned strata (Figure 2.7d). These inclined resistant-strata-bound sets are interpreted to have a slope-dependent relationship.

The similarity in sedimentary texture, tone, and erosional expression between the upper component of the scour fills of CS1, CS2, OS1, and CS3 (Figures 2.6-2.13) and the scour-enclosing strata leads me to interpret a similar depositional process for all. This interpretation is reinforced by the observed continuity of beds and laminations between the upper component of the scour fill and the scour-enclosing stratigraphy. The upper component and the surrounding stratigraphy merge seamlessly together with no observed truncations or interruptions both stratigraphically upwards and laterally beyond the extent of these scour-and-fill structures (Figures 2.6-2.11).

The three-dimensional scour surface morphology is difficult to characterize due to the limited range of outcrop exposure orientations. CS3 (Figure 2.3b and c) occurs on the opposite side of the Chenapau butte to CS1 (Figure 2.3b and c). A reasonable assumption is that these two cliff sections intersect the same three-dimensional scour plane. The width and depth of CS3 is much smaller compared to CS1. I infer that the scour exhibits spatial variability in depth and width, and that the CS3 outcrop intersects the scour at a particularly narrow section.

2.5.2 Class A Scour-and-fill structures with a complex morphology and light-toned infill

2.5.2.1 Observations

Orinoco Scour 1 (OS1)

OS1 occurs to the north of the western face of Orinoco. In profile, it erosively truncates planar strata

of a dark-toned unit (Figure 2.11). Only the southern margin and basal portion of the structure is preserved with a total preserved extent of ~40 m and depth of ~6 m between elevations of -3832 m and -3838 m.

The preserved scour paleo-surface is concave-up but the southern margin shows a complex geometry consisting of a stepped section climbing southwards with a small composite, concave-up section, truncating the underlying planar stratigraphy (Figure 2.11c). The first step (~21 m in width) comprises the lowermost section (Figure 2.12). This section exhibits a planar base which gradually slopes upward into an inclined surface reaching a maximum apparent dip of 21°. The inclined surface gradually flattens out, again becoming parallel with the underlying strata to form a second step (Figure 2.11c). Towards the south, this surface gradually slopes upwards to reach a maximum dip of 11° (spread over the next ~12 m of section). The final section of the scour surface exhibits another small, composite, concave-up section, ~4 m in width.

The scour fill is composed of light-toned, recessive-resistant packages in three components: a lower, middle, and upper fill (Figure 2.11c). The lower fill consists of recessive-resistant packages which drape the lowermost section of the scour surface and onlaps laterally onto the inclined portion of the first step (Figure 2.12). This fill is sharply truncated by a bounding surface that merges laterally into the basal concave surface of the scour-and-fill structure, aligning with the second step. The middle fill drapes the underlying bounding surface and the second step (Figure 2.11c, 12b). It is characterized by prominent bands interspersed with recessive packages, with the number of bands decreasing upward within this component. An additional concave-up bounding surface truncates the middle fill. The upper fill drapes this sharp contact with the middle fill component (Figure 2.11c) and has a distinct light-toned, recessive appearance in which individual strata are not resolved, and which is indistinguishable from the overlying scour-enclosing strata.

Bela Vista Scour (BVS)

A prominent erosional surface is present at the top of the western face of a butte consisting of the diagenetically overprinted unit, Bela Vista. This erosional surface takes the form of a concave-up, scour-shaped feature (Figure 2.13). The surface extends to the outcrop edge and any continuation of this surface beyond the limit of the butte is not preserved. The remnant structure is ~16 m in width and ~1 m deep and occurs at elevations between ~-3842 m and -3843 m. Due to diagenetic alteration, none of the original underlying stratigraphy or sedimentary features can be discerned. This scour is distinct from other scours in Class A as the central portion is smaller, arcuate and not planar.

Despite apparent diagenetic alteration of scour infill, bedding or potential third-order bounding surfaces can be delineated and appear to drape across the scour surface (Figure 2.13). In one location, one of these interpreted third-order bounding surfaces of the scour fill is truncated by a concave-up, simple, second-order bounding surface which appears to converge laterally, gradually merging with the basal concave surface of the structure.

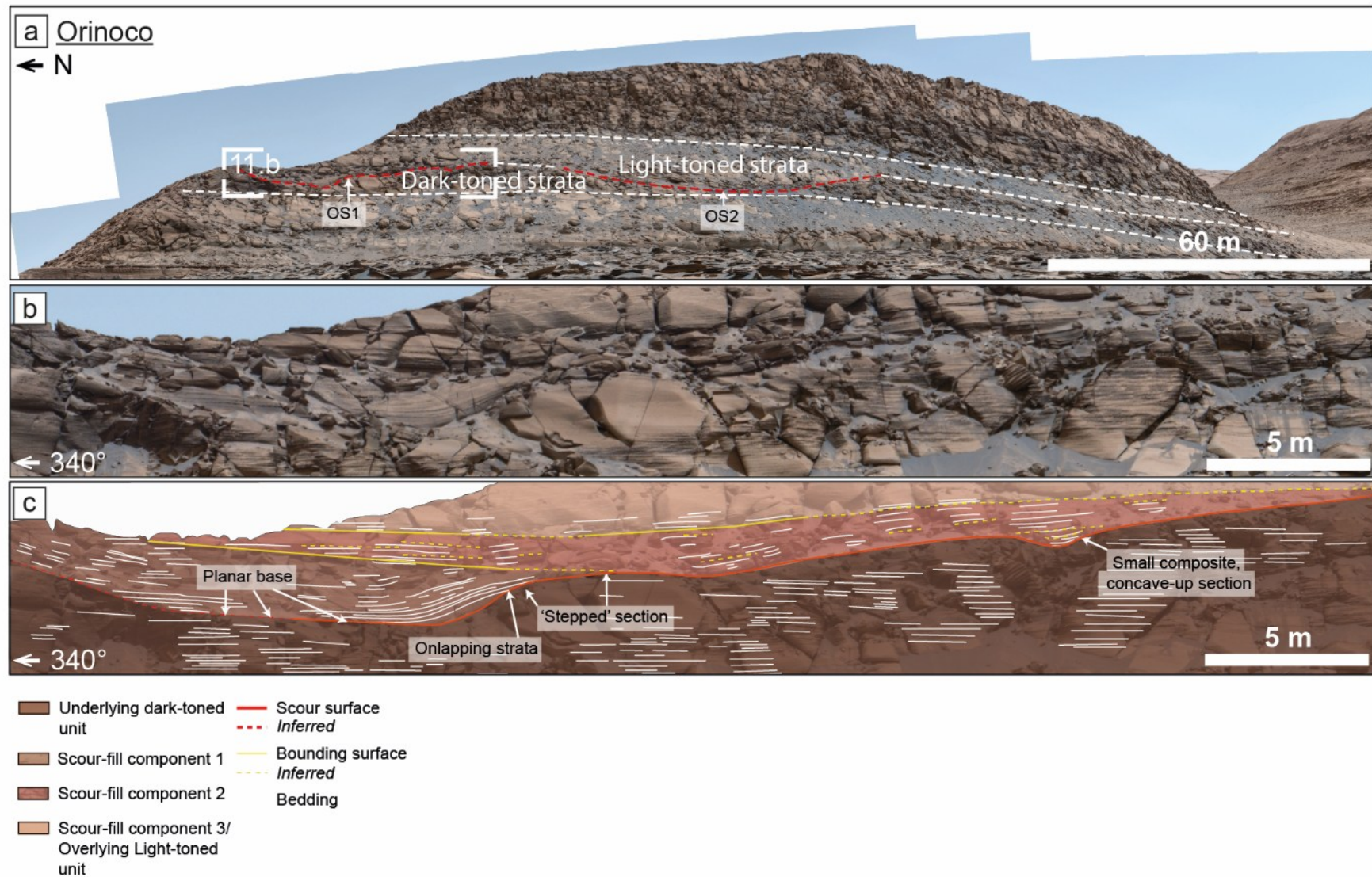


Figure 2.11: Architecture of Orinoco Scour 1 (OS1). (a) Mastcam panorama of the Orinoco butte illustrating the stratigraphic contacts (white dashed lines) and the location of the OS1's erosive base (red dashed lines), north of OS2 (Mastcam mosaic MR_mcam102672 acquired on Sol 3578, at Sol 3576 end-of-drive location). (b) Detail showing OS1 uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop as described in the text. Image credits: NASA/JPL-Caltech/MSSS.

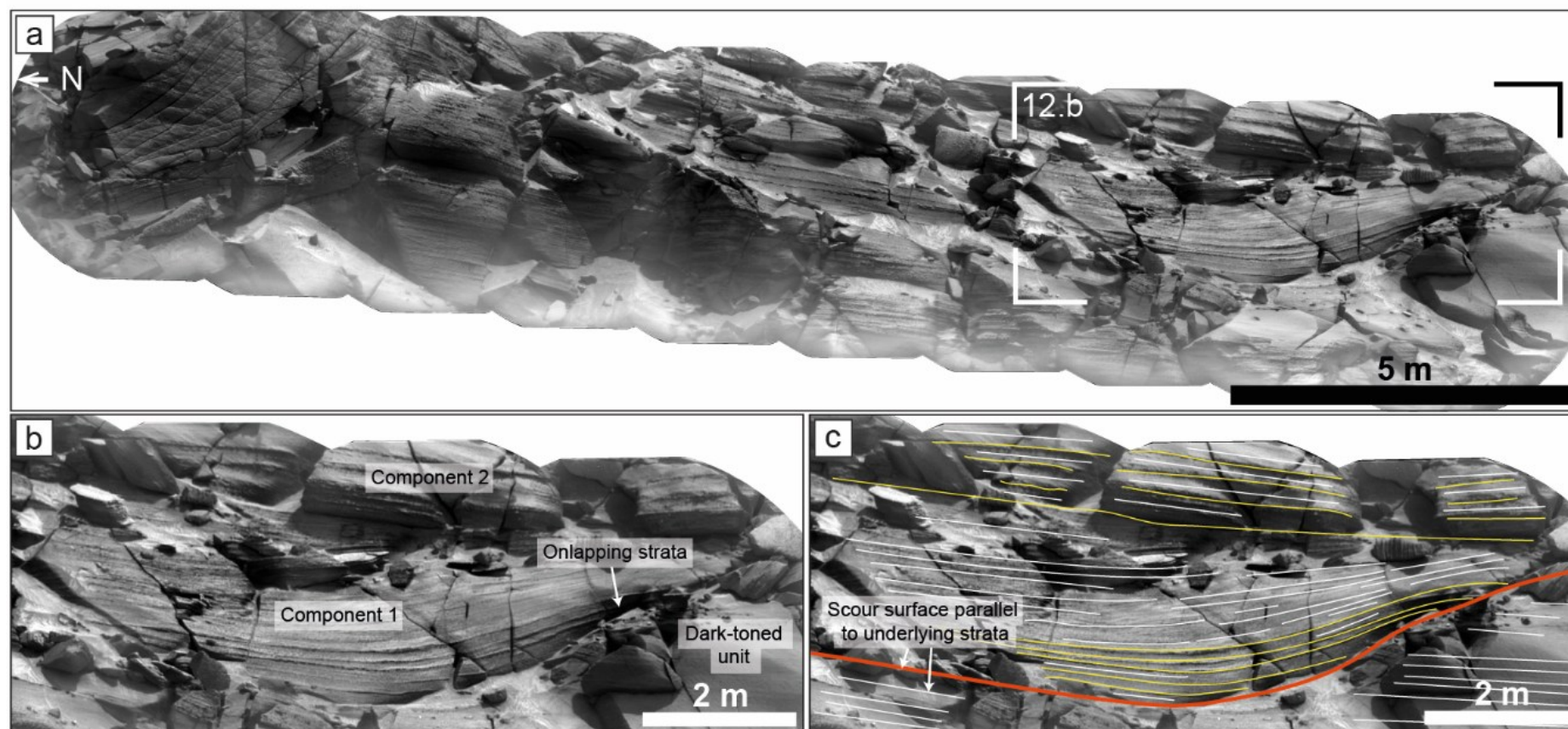


Figure 2.12: Details of Orinoco Scour 1 (OS1). (a) Long Distance Remote Micro-Imager (LD RMI) mosaic of component 1 of OS1 (LD RMI mosaic ccam_04582, acquired on Sol 3582, at Sol 3579 end-of-drive location). (b) Detail of showing how the infill of component 1 onlaps onto the scour surface and is truncated by component 2. (c) Interpretation showing the scour surface (red), bounding surfaces (yellow), and bedding (white). Image credits: NASA/JPL/LANL/CNES/IRAP.

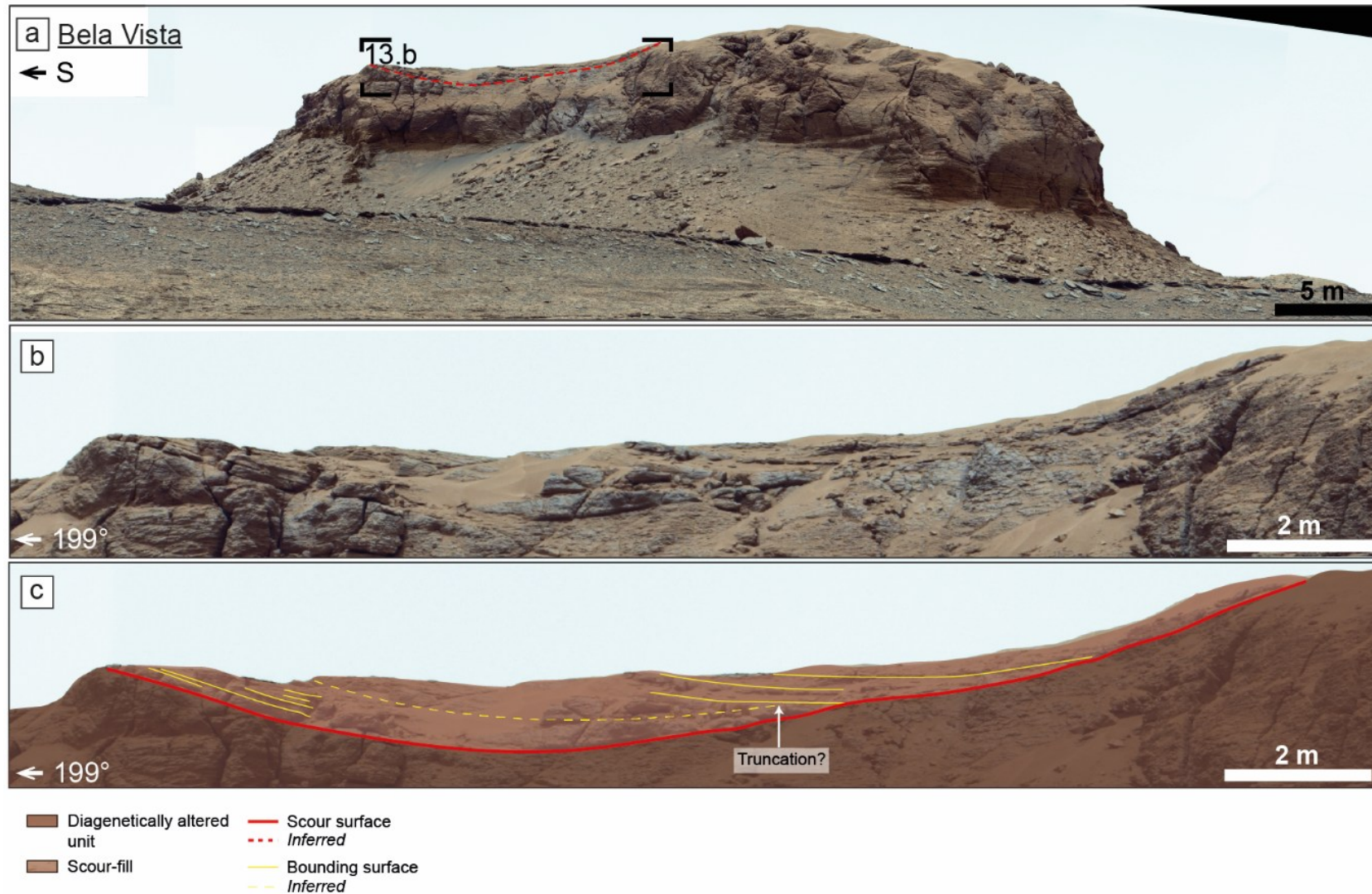


Figure 2.13: Architecture of Bela Vista Scour (BVS). (a) Mastcam panorama of the Bela Vista butte illustrating the stratigraphic contacts (white dashed lines) and the locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mcam102968, acquired on Sol 3635, at Sol 3633 end-of-drive location). (b) Detail showing BVS uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop. Image credits: NASA/JPL-Caltech/MSSS.

2.5.2.2 Interpretation

I interpret OS1 and BVS as a complex of stacked scour-and-fill structures formed by amalgamation of simple, smaller scour-and-fill structures. This interpretation is supported by the observation of distinct erosional surfaces or second-order bounding surfaces truncating the scour infill (Figure 2.11c and Figure 2.13c) that appear to merge with the underlying basal scour-surface. In OS1, these surfaces are also associated with each of the three “stepped” concave sections of the scour surface (Figure 2.11) and are thus interpreted to form the shape of three stacked scour-and-fill structures. When OS1 and BVS are divided into individual stacked scour components, the scour bases themselves are simple, concave-up, broad, and OS1 even contains a planar basal section. These characteristics are similar to the morphology of the basal scour surfaces exhibited in Section 2.5.1.

I also note a similar scour fill conforming relation in OS1 and BVS to the scours in Section 2.5.1. The infills of OS1 and BVS also conform to the underlying scour surface or erosional bounding surfaces (Figure 2.12). Corroborating this, I also observe an upper component in OS1 which shares a similar light-tone and texture to the upper component infill within scour-and-fill structures in Section 2.5.1. Therefore, our interpretation leads me to the conclusion that the infill processes for the stacked scours of OS1 are also comparable to the individual scours in Section 2.5.1. Hence, I infer that OS1 and BVS may share similarities with the erosional processes of the scours in Section 2.5.1, albeit with some added complexities. OS1 also occurs in the same stratigraphic interval and within the same ~9 m topographic elevation as CS1, CS2, ES1, ES2, CS3, and OS2. Based on these factors, I suggest that OS1 may have formed during the same episode as the scour-and-fill structures identified in Section 2.5.1.1.

2.5.3 Class A scour-and-fill structures that have been exhumed and contain minimal to no preserved infill

2.5.3.1 Observations

Exhumed Scour 1 (ES1)

ES1 occurs within a distinct ledge present along the southwestern flank of Marker Band valley (Figure 2.14). It is characterized by a broad, concave-up gap carved into the bedrock, forming a curved surface that appears to incise the underlying stratigraphy (Figure 2.14b). This curved surface is symmetric and uniform, measuring approximately 40 m in width and 1 m in amplitude, and occurs at elevations ranging from ~3834 m to ~3832 m (Figure 2.14). The margins exhibit apparent dips of 12–13° relative to the underlying strata, while the central portion of the surface is planar and parallel to the underlying stratigraphy.

Recessed behind the ledge and overlying the interpreted northern margin of the scour is a thin band of preserved bedding inclined at a different angle to strata underlying the scour (Figure 2.14c). The preserved bedding within this band appears conformable to the scour surface along the northern margin

and scour base. No contact is visible, but this may be due to dust coverage, terrain occlusion, and strong foreshortening due to the viewing geometry.

Exhumed Scour 2 (ES2)

Directly adjacent to ES1 along the same ledge is a second exhumed scour forming a broad, concave-up gap eroded into bedrock (Figure 2.15). This feature has a width of ~50 m and amplitude of ~3 m and occurs between ~-3841 m and -3838 m elevation. The scour surface is broad and symmetrical, with preserved scour margins showing apparent dips of ~16° relative to the underlying planar strata. In one location, strata can be traced for ~2 m along a sharp contact with the northern scour margin (Figure 2.15d). These strata appear to drape the scour's paleo-surface (Figure 2.15e).

Eroded Remnants of an Exhumed Scour 3 (ES3)

Recessed to the west behind ES1 and ES2 is a second stratigraphically overlying ledge. The ledge is inclined to the north, flattening out along an asymptotic base (Figure 2.16). This inclined morphology strongly resembles a scour margin, thus I infer this feature represents a scour surface. This eroded remnant of a scour surface occurs at elevations between -3819 m and -3823 m. It has a preserved width of ~19 m and an amplitude of ~4 m. No infilling strata are observed.

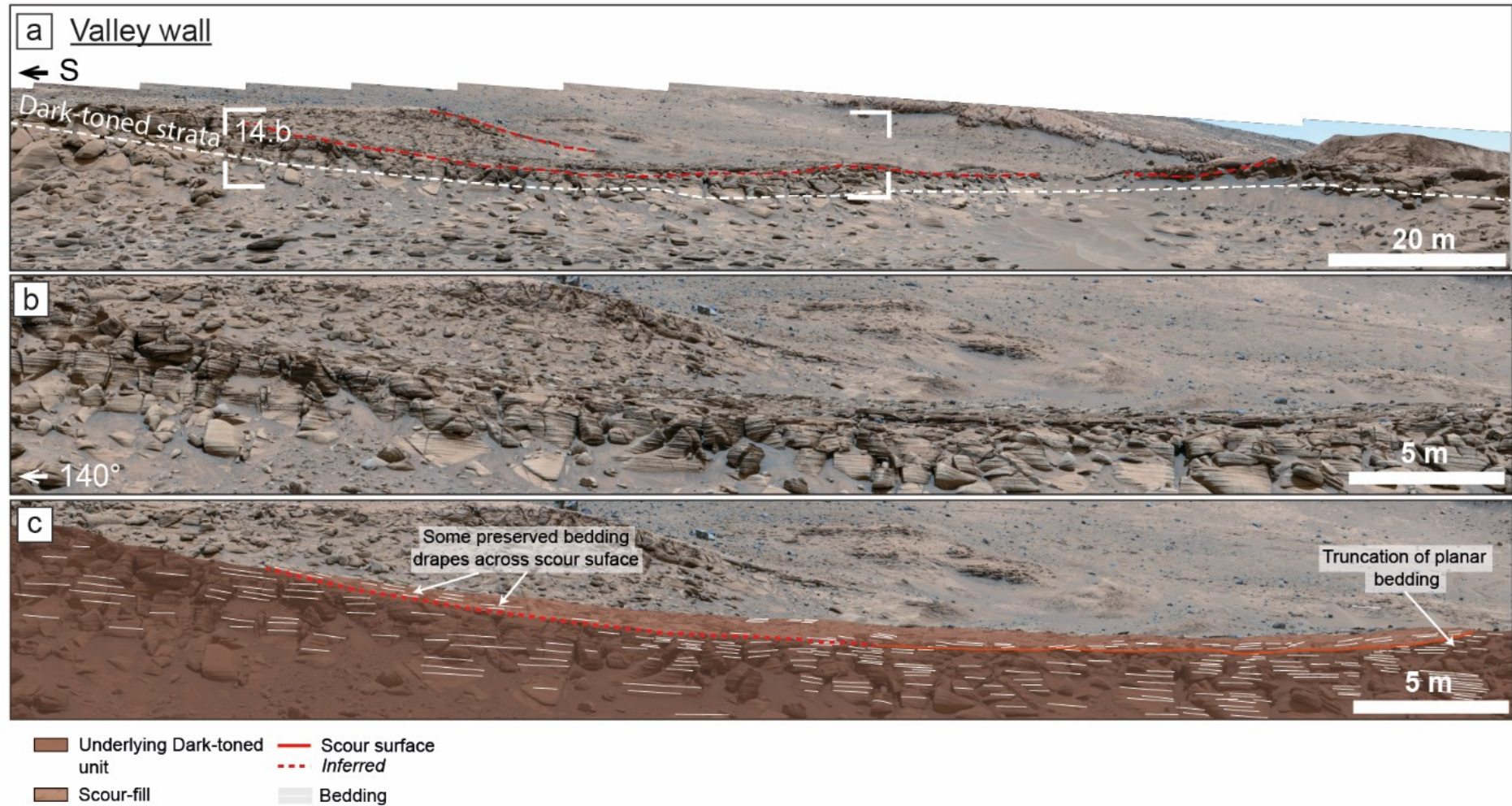


Figure 2.14: Architecture of Exhumed Scour 1 (ES1). (a) Mastcam panorama of the westernmost wall of Marker Band valley illustrating the stratigraphic contacts (white dashed lines) and the locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mcam103501 acquired on Sol 3730, at Sol X end-of-drive location). (b) Detail showing ES1 uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop. Image credits: NASA/JPL-Caltech/MSSS.

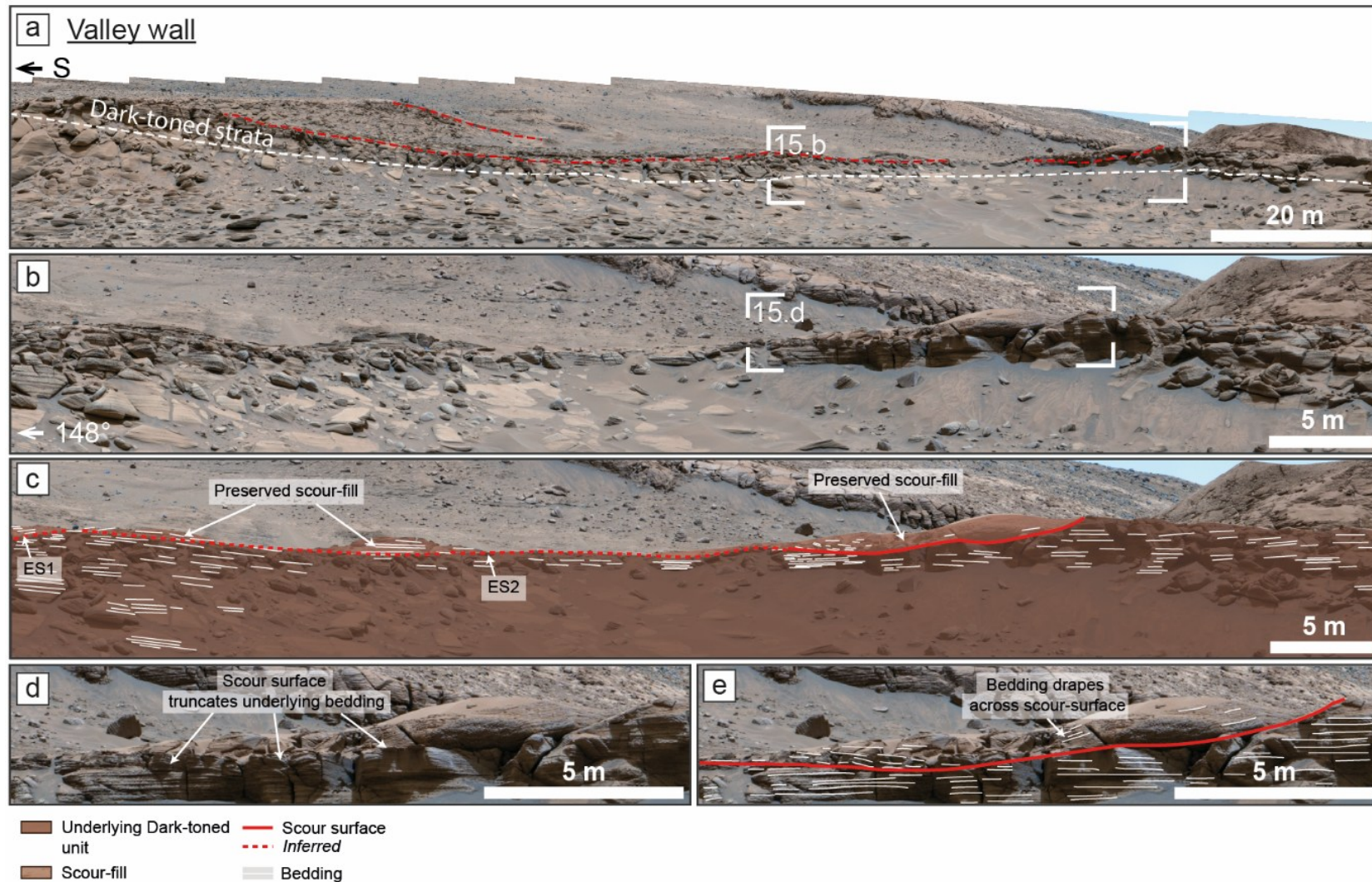


Figure 2.15: Architecture of Exhumed Scour 2 (ES2). (a) Mastcam panorama of the westernmost wall of Marker Band valley illustrating the stratigraphic contacts (white dashed lines) and the locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mcaml03501 acquired on Sol 3730, at Sol 3728 end-of-drive location). (b) Detail showing ES2 uninterpreted. (c) Interpretation showing the scour boundaries, bounding surfaces, and stratigraphy mapped in the outcrop as described in the text. (d) Detail showing how the scour surface is observed truncating the underlying bedding and is overlain by partially preserved scour fill that drapes across the scour surface. (e) Interpretation showing the scour's erosive surface (red) and bedding (white). Image credits: NASA/JPL-Caltech/MSSS.

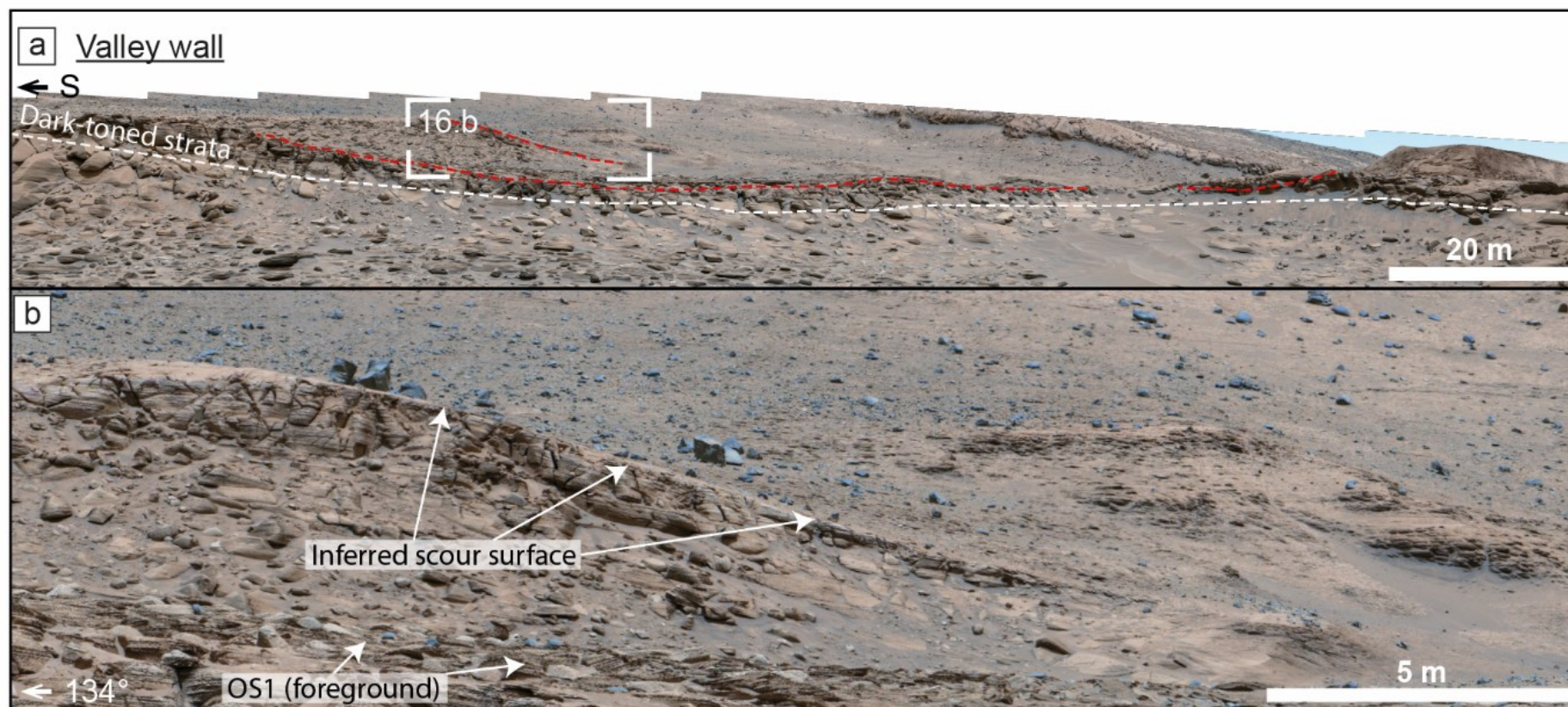


Figure 2.16: Architecture of Exhumed Scour 3 (ES3). (a) Mastcam panorama of the westernmost wall of Marker Band valley illustrating the stratigraphic contacts (white dashed lines) and the locations of the scour-and-fill structures' erosive bases (red dashed lines) (Mastcam mosaic MR_mc103501 acquired on Sol 3730, at Sol X end-of-drive location). (b) Detail of showing the partially preserved, inferred erosive surface of ES3. Image credits: NASA/JPL-Caltech/MSSS.

2.5.3.2 Interpretation

ES1, ES2, and ES3 all exhibit a strong morphological similarity to the scours in Section 2.5.1, thus I infer a similar formation process. These scours all show a similar range of widths (~40 m to ~62 m) and amplitudes (~1 m to ~4 m) (Table 2.1). The simple, symmetric, concave-up morphologies are also consistent. The maximum apparent dip of the scour margins of ES1, ES2, and ES3 varies around 12-16° which is comparable to the dips of the scour margins in Section 2.5.1. ES1 and ES2 also exhibit a planar central scour base. Moreover, ES1 and ES2 occur within the same ~9 metres of elevation as CS1, CS2, CS3, OS1, and OS2 (Table 2.1). For this reason - and because the exhumed scours truncate into the same stratigraphic interval at similar elevations - I also interpret that scours ES1 and ES2 formed during the same temporal window as CS1, CS2, CES, and OS1.

Where bedding overlies the scour surface in ES1 and ES2, stratification conforms to the underlying scour, mirroring the scour infill characteristics of the scours found in Section 2.5.1 and 5.2. Given the draping relationship, similar distinctions in erosion resistance between scour fill and the underlying strata, and as they occur at the same stratigraphic interval, I interpret the depositional processes of ES1 and ES2 to be similar to the scour-and-fill structures in sections 2.5.1 and 2.5.2.

Exhumation of these scours may have been caused by differences in erosional resistance between scour infill and the underlying strata. The infill was likely less resistant to erosion relative to the underlying stratigraphy (e.g. Edgett and Sarkar, 2021) and thus more easily exhumed by weathering processes, leaving the scour-shaped gaps in the underlying dark-toned strata.

2.5.4 Class B Scour-and-fill structures

2.5.4.1 Observations

Canta Detour Scour (CDS)

A concave-up scour-and-fill structure containing indurated, dark-toned infill (Figure 2.17) was observed eroded into light-toned, diagenetically overprinted unit along the western flank of Marker Band valley. This structure is ~7 m in width, <0.5 m deep, and occurs at ~3850 m elevation. While the eastern scour margin is visible, the western margin is self-occluded by outcrop viewing geometry. The lower scour surface forms a singular concave-up scoop along a sharp contact with underlying strata. An additional composite feature is present: an inflection point on the eastern scour margin where scour morphology transitions from concave to convex as it pinches out as a wing-shaped structure (Figure 2.17c).

The dark-toned infill consists of amalgamated bedsets consisting of centimetre-scale tabular cross-stratification dipping to the north-west (Figure 2.17d, e). These sets are consistently truncated by overlying tabular sets. The scour infill contact of with the overlying stratigraphy is sharp. The infill appears to be more indurated than the structure-enclosing light-toned recessive strata.

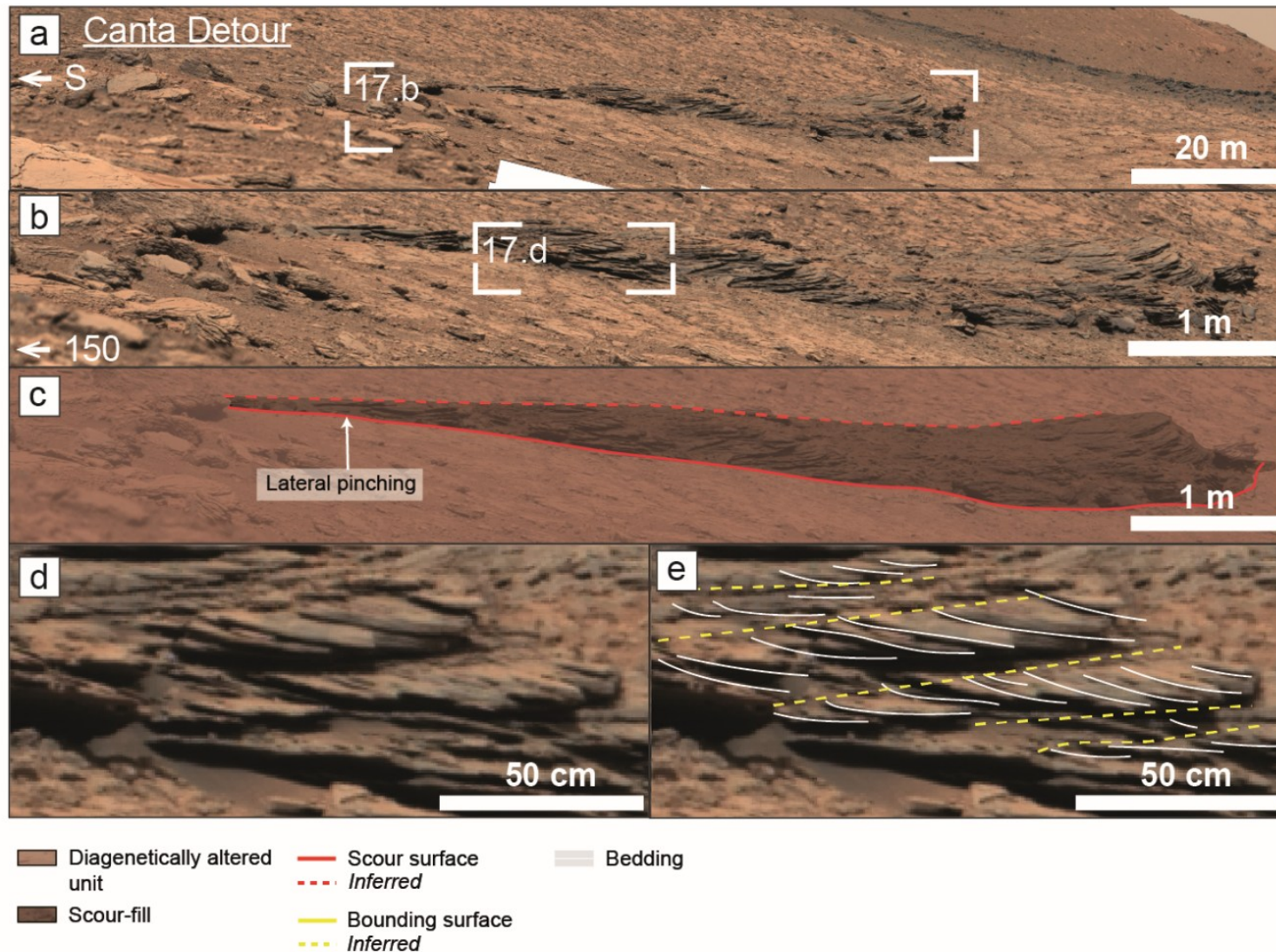


Figure 2.17: Architecture of Canta Detour Scour (CDS). (a) Mastcam panorama of part of the Canta Detour providing context for the scour-and-fill structures (Mastcam mosaic MR_mcam103090 acquired on Sol 3653, at Sol 3651 end-of-drive location). (b) Detail showing CDS uninterpreted. (c) Interpretation showing the scour boundaries mapped in the outcrop as described in the text. (d) Detail showing the scour fill. (e) Interpretation showing the cm-scale cross-bedding as described in the text. Image credits: NASA/JPL-Caltech/MSSS.

2.5.4.2 Interpretation

Class B is the most distinctive of the scour-and-fill classes in terms of both morphology and infill characteristics. The strongest difference between CDS, the sole scour in class B, and class A is scale. CDS is ~50 to ~90% smaller than the other scour-and-fill structures described in this study (Table 2.1). CDS is the sole scour exhibiting a pinching out wing-shape morphology. The infill of CDS is dark-toned, erosionally resistant, and characterized by multistorey, erosive decimetre-scale sets. This contrasts with the large-scale, light-toned recessive bedsets conforming to the scour surface observed in class A. These architectural distinctions suggest class B's infill differs from the other scour-and-fill structures and the enclosing aeolian sandsheets. The distinction in erosion resistance and tone suggest that the class B's infill may also compositionally differ from the enclosing aeolian sandsheets. Due to these distinctions in size, morphology, and infill characteristics, I interpret that the processes that scoured and infilled class B were different from the other scours.

2.6. Hypotheses for formation of scour-and-fill structures

Ten scour-and-fill structures, and their associated infill where present, have been identified within the Chenapau member of the Mirador formation exposed in Marker Band valley. These structures exhibit a range of morphologies and may contain light-toned and dark-toned sedimentary infill. Several scour structures in which the infill has been exhumed were also identified (Figure 2.18). Class A scour-and-fill structures are interpreted to have formed through the same erosional and depositional processes due to similarities in their size, morphology, and scour infill characteristics. The class B scour-and-fill structure is distinct from class A and is discussed separately.

The host stratigraphy facies consisting of planar, uniform-in-thickness, and laterally extensive planar laminae are characteristic of formation by the migration of aeolian wind-ripples (e.g. Kocurek and Nielson, 1986) (Section 2.4.1; 4.2). The observation that this facies occurs in planar parallel sub-horizontal bedsets suggests that this facies represents deposition in an aeolian sandsheet environment (Kocurek and Nielson, 1986). On this basis, I infer that the scour-and-fill structures were formed in an overall aeolian sandsheet setting. Scouring from fluvial processes, bedform migration, and aeolian deflation are all capable of forming scour-and-fill structures in the sedimentary record (e.g. Blakey, 1988; Kocurek and Day, 2018; Priddy and Clarke, 2020). Here, I evaluate and discuss these processes.

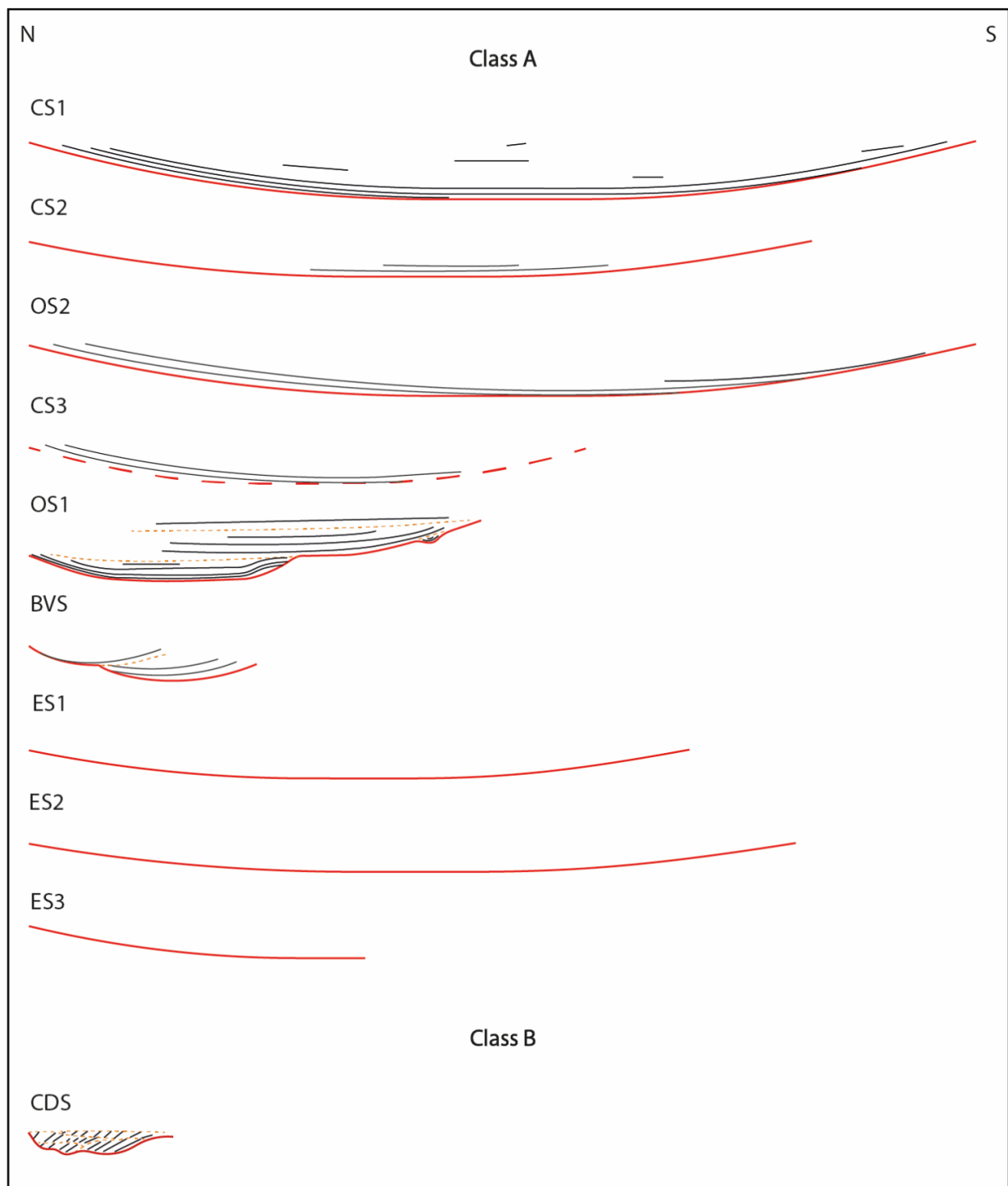


Figure 2.18: Reconstructed geometry of each of the scour-and-fill structures to scale. Scour surface is in red, strata is in black and bounding surfaces (when appropriate) are in orange. Dashed lines show inferred geometries. Note the size difference and scour fill architecture between classes A and B.

2.6.1. Hypothesis 1: A fluvial mechanism for scour-and-fill structures

The distinctive architectural features and facies associations commonly observed in channel scour cross-sections resulting from fluvial erosion are well documented (e.g. Gibling, 2006). In the Chenapau member, the scours exhibit width-to-depth (W/T) ratios of ~15. Fluvial scours with similar width-to-depth ratios are inconsistent with typical channels and instead are commonly associated with shallow channels and sheet-form fluvial deposits found in aeolian environments (Figure 17a; e.g. Gibling, 2006; Priddy and Clarke, 2020). Similar sized, broad, concave-up scours have previously been inferred by Rapin et al. (2021) in the layered sulfate unit in Aeolis Mons from ~3 to 6 km away. These observations led Rapin et al. (2021) to interpret these scours as fluvial in origin, however, this interpretation remains open as the facies details could not be fully resolved at such observational distances.

There are a number of expected infill characteristics associated with fluvial scouring in an aeolian environment that are inconsistent with scour class A. Scour-and-fill structures in class A do not contain evidence for lag components with pebbles >1 cm in diameter—which would likely be visible at most of the scour image resolutions (Table II.5, Appendix II) – and rip up clasts which are often associated with episodic fluvial flooding in aeolian environments (e.g. Banham and Mountney, 2014; Cain and Mountney, 2009; Priddy and Clarke, 2020). The scour infill does not show evidence for centimetre-scale current ripple lamination nor tens of centimetre-scale cross-bedding which would be evidence for migration of dunes within a fluvial channel (e.g. Banham and Mountney, 2014; Cain and Mountney, 2009; Gibling, 2006). Instead, the similarity and the observed continuity of the scour fill of Class A with the overlying light-toned, scour-enclosing strata indicates a shared depositional process. With the scour-enclosing strata interpreted to be from the migration of aeolian wind-ripples (Section 2.4.2), I propose an aeolian interpretation for the scour fill.

The morphology of the scour surfaces in class A are inconsistent with fluvial channel morphology. The scours in class A have no clear inflection points or pinchout into a wing-shape at their lateral extents, a key feature in fluvial architecture caused by processes such as overbank flooding (e.g. Buffington and Montgomery, 2022; Gibling, 2006). The base of the scours are also predominantly smooth with none of the irregularities associated with varying flow rates and resistance found in channels along cross-section (e.g. Charlton, 2007; Wohl, 1998) common in shallow channels and sheet-like deposits in aeolian environments (e.g. Ferronato et al., 2019; Heme de Lacotte and Mountney, 2022; Jaeger et al., 2017; Langford and Chan, 1989; Scherer and Lavina, 2005). The limbs of the scour surfaces in the Chenapau member are also shallow with apparent dips of ~10-20°, inconsistent with the common “steep” sides recorded by Gibling (2006) for paleochannels in aeolian settings. As the scours in class A are inconsistent with characteristics typical from fluvial scouring, I infer that they likely formed by a different erosional process.

The distinct morphology and infill characteristics of scour-and-fill structure class B, however, are consistent with fluvial scouring in an aeolian environment. The inferred asymmetry of the central body

and isolation to any other similar scour forms may indicate deposition in a channelized, single-threaded river (e.g. Engel and Rhoads, 2012; Gibling, 2006; Thornbury, 1954). The pinching out seen in the southern flank of CDS is characteristic of channel wings formed through aqueous scouring in channels from overbank flooding (e.g. Buffington and Montgomery, 2022; Gibling, 2006). The width-to-thickness ratio of ~ 14 observed in class B supports the interpretation of a small, shallow fluvial channel formed by fluvial processes within aeolian environments (e.g. Gibling, 2006). The distinctive, amalgamated bedsets of cm-scale tabular cross-stratification are consistent with infilling by fluvial dune migration (e.g. Gibling, 2006). The darker tone and induration of the scour fill in class B, compared to the surrounding stratigraphy and scours, suggest a distinct sediment infill or even provenance relative to the adjacent strata. Class B also aligns with several metre-scale, dark-toned, indurated lenses interstratified with aeolian dunes in the Contigo member interpreted to be aqueous lenses (Figure 2; Caravaca et al., 2022a; Gupta et al., 2022). While class B represents an aqueous depositional environment, the distinction in scour architecture to Class A implies different formative mechanisms and sedimentary processes for the rest of the scours.

2.6.2. Hypothesis 2: Aeolian Bedform Migration as a Mechanism for Scour-and-fill Structures

One plausible hypothesis for the formation of the scour-and-fill structures is by aeolian dune migration (Figure 17b; e.g. Cardenas et al., 2022; Kocurek and Havholm, 1993; Mesquita et al., 2021; Mountney and Thompson, 2002; Swanson et al., 2019). In this mechanism, the erosion of the scour surface is created by migration of troughs leeward of a migrating dune which gets progressively infilled as the bedform migrates into the scour. Dune deposits can occur as concave-up scour-and-fill structures when the overall angle of bedform climb - determined by the balance between erosion and deposition - is low but fluctuates over shorter time scales (Figure 17b; Abrantes et al., 2020; Kocurek and Day, 2018; Kocurek and Havholm, 1993; Mesquita et al., 2021; Mountney and Thompson, 2002). Alternatively, migration of three-dimensional dunes with laterally variable scour depths can result in especially deep troughs being preferentially preserved as scour-and-fill structures (Cardenas et al., 2019; Swanson et al., 2019). However, I find that class A scour-and-fill structures are distinct from those predicted by bedform migration. Scour-and-fill structures from bedform migration likely contain evidence of cross-bedding - tangential, tabular (Cardenas et al., 2019; Hasiotis et al., 2021; Kocurek and Day, 2018; Mesquita et al., 2021; Mountney and Thompson, 2002) or trough-shaped (Abrantes et al., 2020) - along with clear truncation of the scour fill by overlying dune or interdune deposits. The infill within class A scour-and-fill structures, however, although superficially similar to trough cross-bedding, lacks these clear truncation surfaces (Figure 2.19b), suggesting that class A does not contain a discrete, trough cross-bedded set. Instead, class A scour-and-fill structures appear overfilled with an infill that seamlessly merges with the overlying and surrounding stratigraphy (Section 2.5.1.2). The

continuity in sedimentary texture, tone, and erosional patterns between the upper component of the scour fill - where preserved - and the surrounding sandsheet facies (Section 2.4) supports an interpretation of sand-sheet migration into a pre-existing topographic low. Additionally, class A scour-and-fill structures are laterally isolated, whereas scour-and-fill structures formed by aeolian bedform migration would likely result from a field of bedforms. These features often display lateral truncation between adjacent scours (e.g. Cardenas et al., 2019), a characteristic that is absent in class A scour-and-fill structures. Therefore, I argue that it is unlikely the scours in the Chenapau member were caused by bedform migration.

2.6.3. Hypothesis 3: Scour formation by aeolian blowouts

An alternative hypothesis for formation of the scour-and-fill structures is by aeolian blowout deflation (Figure 17c; Blakey, 1988; Gradzinski, 1992; Hesp, 2002; Warner et al., 2022). Blowouts are depressions or hollows formed through localized removal of sediment by wind erosion of a stabilized surface in an arid to semi-arid setting (Hesp, 2002; Hesp, 1996). Although vegetation often acts as the stabilizing agent on Earth, blowouts also occur in regions with little-to-no observed vegetation, where other surface stabilizing mechanisms such as cementation, lag deposits, and subsurface ice dominate (Hesp, 2002; Hesp and Hyde, 1996). The concave-up scours in the Chenapau member resemble 2D cross sections through saucer-shaped blowout structures (e.g. Hesp and Walker, 2013). The widths and thicknesses of the scours have comparable diameters and depths to blowout structures observed on Earth (e.g. Baird et al., 2021; Hugenholtz and Wolfe, 2006). Recognizing blowout depressions in the rock record can be challenging due to a lack of outcrop continuity (e.g. Blakey, 1988; Ferronato et al., 2019; Gradzinski, 1992), yet these features are commonly observed in Holocene and present-day aeolian geomorphology (e.g. Hesp, 2002) and have been extensively studied. I can compare terrestrial blowouts to the Gale scour-and-fill structures by drawing on studies of the geomorphology and settings of recently active blowouts (e.g. Abhar et al., 2015; Hesp, 2002; Hugenholtz and Wolfe, 2009), their facies associations and architecture from ground penetrating radar (GPR) surveys (e.g. Jewell et al., 2017; Pedersen and Clemmensen, 2005; Warner et al., 2022) and examples from the rock record (e.g. Blakey, 1988; Ferronato et al., 2019; Gradzinski, 1992).

Blowout Erosional Deflation Processes

The broad, consistent, symmetric, concave-up morphologies and flat bases exhibited by the scour-and-fill structures are characteristic of processes that form blowouts. The distinctive regular, symmetric morphologies of the “saucer” and “bowl” blowout types is a product of a morphology and wind-flow feedback loop, where a wind-steering effect induced by the rounded blowout walls creates a re-circulation cell (Hesp and Walker, 2012; Hugenholtz and Wolfe, 2009). This process, while removing most sediment from the blowout, also redistributes sediment from the deflation around the “saucer,” maintaining the characteristic, symmetric, consistently semi-circular depression and gradually evening

out any irregularities in shape (Hesp, 2002; Hesp et al., 2017; Hugenholtz and Wolfe, 2009). The flat bases exhibited by the Chenapau member scours are similar to flat surfaces which form in blowouts when wind deflation, and thus the depth of a scour, meets a stabilized surface lower down in the stratigraphy such as a hard, indurated layer or the water table (e.g. Hesp and Walker, 2013). In these instances, Hesp and Walker (2012) found that a blowout would expand laterally, rather than vertically downward, forming a flat base. These broad, consistent, concave-up and planar base features are characteristic of blowout cross-sections identified within the rock record (Ferronato et al., 2019; Gradzinski, 1992) and from GPR surveys (e.g. Jewell et al., 2017; Pedersen and Clemmensen, 2005; Warner et al., 2022). For these reasons, I find that the Chenapau member scour-and-fill structures have morphologies consistent with blowout scour formation.

Blowout Infilling processes

The onlapping, downlapping and conformable relationships of the inclined bedsets from the lower component of the scour fills indicate that infill occurred following scour formation. Multiple field examples describe blowout infill with inclined bedsets conformable with the scour base and onlapping/downlapping relationships in both modern ground penetrating radar surveys (e.g. Neal and Roberts, 2001; Pedersen and Clemmensen, 2005; Warner et al., 2022) and in the rock record (e.g. Ferronato et al., 2019; Gradzinski, 1992; Mountney, 2006b). In the latter, Ferronato et al. (2019) and Gradzinski (1992) observed wind-ripple, grain-flow, and grain-fall strata within inclined cross-strata which infilled blowouts which led these authors to an interpretation that the inclined cross-sets were formed by dunes - possibly parabolic or crescentic - building into and migrating across the blowout. Warner et al. (2022) proposed a similar dune migration process for cross-stratified blowout-infills from GPR studies of Holocene blowouts. The upper range of apparent dips of the lower fill component of the scour-and-fill structures are similar to the apparent dips identified within the GPR cross-sections of the blowout infill in Warner et al. (2022).

While the resolution of our data is limited and I cannot resolve sedimentary textures indicative of grain-flow or grain-fall strata like Ferronato et al. (2019) and Gradzinski (1992), I did observe bands separating the downlapping and onlapping sets. These bands were comparatively more erosionally resistant than the more recessive light-toned sets and showed some truncation of underlying sets, evidence of reactivation. The observed differences in the erosional resistance is likely a result of the complex interplay between grain packing, grain size, and cementation (Banham et al., 2024a; Besly et al., 2018). I speculate that the differences in resistance could represent changes in sedimentary texture from slope depositional processes found in migrating dunes. However, unlike the inclined cross-sets interpreted as dune cross-stratification from the rock record (e.g. Ferronato et al., 2019; Gradzinski, 1992; Mountney, 2006b) and GPR surveys (Warner et al., 2022), the lower component bedsets identified in our data make up less than 25% of the scour fill component, do not show truncation from the scour-enclosing strata, and drape across the entirety of the underlying scour surface rather than

exhibit lateral migration at dips parallel to one of the scour limbs.

The light-toned, recessive, planar upper scour fill component which I interpret as wind-ripple laminae from a sand-sheet environment (Section 2.5.1.2) holds some similarities to the fill of blowouts in the rock record. Both Ferronato et al. (2019) and Gradzinski (1992) identified blowouts with “an over-filled appearance” exhibiting a transition from inclined cross-stratification to more planar strata across the top of the scour margin. Based on these observations, Ferronato et al. (2019) interpreted the aeolian sand sheets to be laterally continuous with aeolian dunes that filled the blowouts. Similar to Ferronato et al. (2019), I find that sand sheets are laterally continuous with the inclined cosets infilling the blowouts. I find that these resistant bands defining our lower scour fill component are lost up section and the scours are filled with the upper component comprised of sand-sheet strata making up the rest of the 75% of the scour fill.

Blowout amalgamation processes

OS1 is interpreted to represent a multistorey body formed by three stacked scour-and-fill structures (Section 2.5.3.2) which form three distinct steps in the southern scour limb as the scours vary in thickness and width (Figs. 15, 17). Similar stepped patterns are found in active blowouts and found to be caused by blowout amalgamation (Smith et al. (2017). Saucer and bowl blowout amalgamation is a common process in blowout development (e.g. Abhar et al., 2015; Smith et al., 2017). Abhar et al. (2015) used satellite data to show complex blowout bodies formed from the amalgamation of individual blowouts. Smith et al. (2017) observed active blowout amalgamation where a saucer blowout was captured by an adjacent bowl blowout over a period of three years. Amalgamated blowouts have also been identified in the rock record Ferronato et al. (2019), and through GPR studies (Pedersen and Clemmensen, 2005). Our observations of multistorey blowouts are consistent with formation by amalgamation of laterally adjacent blowout structures.

Scour-enclosing environment and blowout initiation

Blowouts can occur in a multitude of arid to semi-arid landscapes including coastal settings (Abhar et al., 2015; Hesp, 2002; Warner et al., 2022), lakesides (Neal and Roberts, 2001), playas (Allen and Anderson, 2000; Goudie, 2023), inland deserts (Hesp, 1996), and even in the sub-arctic (Kayhko, 2007). Saucer-shaped blowouts are often associated with sandsheet environments in both modern (Hesp, 2002; Hugenholtz et al., 2007; Kocurek and Nielson, 1986) and ancient examples (e.g. Ferronato et al., 2019; Gradzinski, 1992). Sandsheets are considered to form in specific aeolian environments where dune development is inhibited by processes such as near surface cementation, a high water table, armoring of the surface by lag deposits or vegetation which all raise the threshold velocity of sand entrainment (Kocurek and Nielson, 1986). Here, evidence of episodic surface water identified throughout the Mirador formation (Caravaca et al., 2022a; Gupta et al., 2023a) and within the study area supports either a high water table or near-surface cementation as the process supporting sand-sheet development (Section 2.4.2). The environments conducive to sand-sheet formation also act to inhibit wind

entrainment of sand (Kocurek and Nielson, 1986) and thus could act as a stabilizing surface, meeting one of the conditions for blowout formation.

The cause of initiation and development of the scour-and-fill structures within the Chenapau member is unclear. Wind has been observed (Hesp and Hyde, 1996) and modelled (Barchyn and Hugenholtz, 2013) to generate small-scale, localized, negative topographic change during a single event on a stabilized surface. Subsequent wind erosion can then exploit this localized negative topography, sustain deflation, and lead to formation of large-scale blowout structures (Hugenholtz and Wolfe, 2006; Smyth et al., 2019; Smyth et al., 2013). It is conceivable that a similar process may have been responsible for the scour-and-fill structures within the Chenapau member.

2.6.4 Evolutionary model for scour-and-fill structure formation

I present the following scenario as a possible sequence of events in the erosional and depositional history of the scours:

- 1) Previously established sandsheets were stabilized or partially stabilized by either early syndepositional near-surface cementation or water table fluctuations.
- 2) Strong winds cause partial deflation creating an uneven topography and small proto-blowouts.
- 3) Sustained, high-velocity winds with low sediment-saturation caused reworking and continued deflation of these blowouts to form the broad, saucer shapes.
- 4) Amalgamation of adjacent blowouts.
- 5) Migration of sand-sheets over and into the blowouts.
- 6) Covering of blowout structures and blowout infill by sandsheet migration.

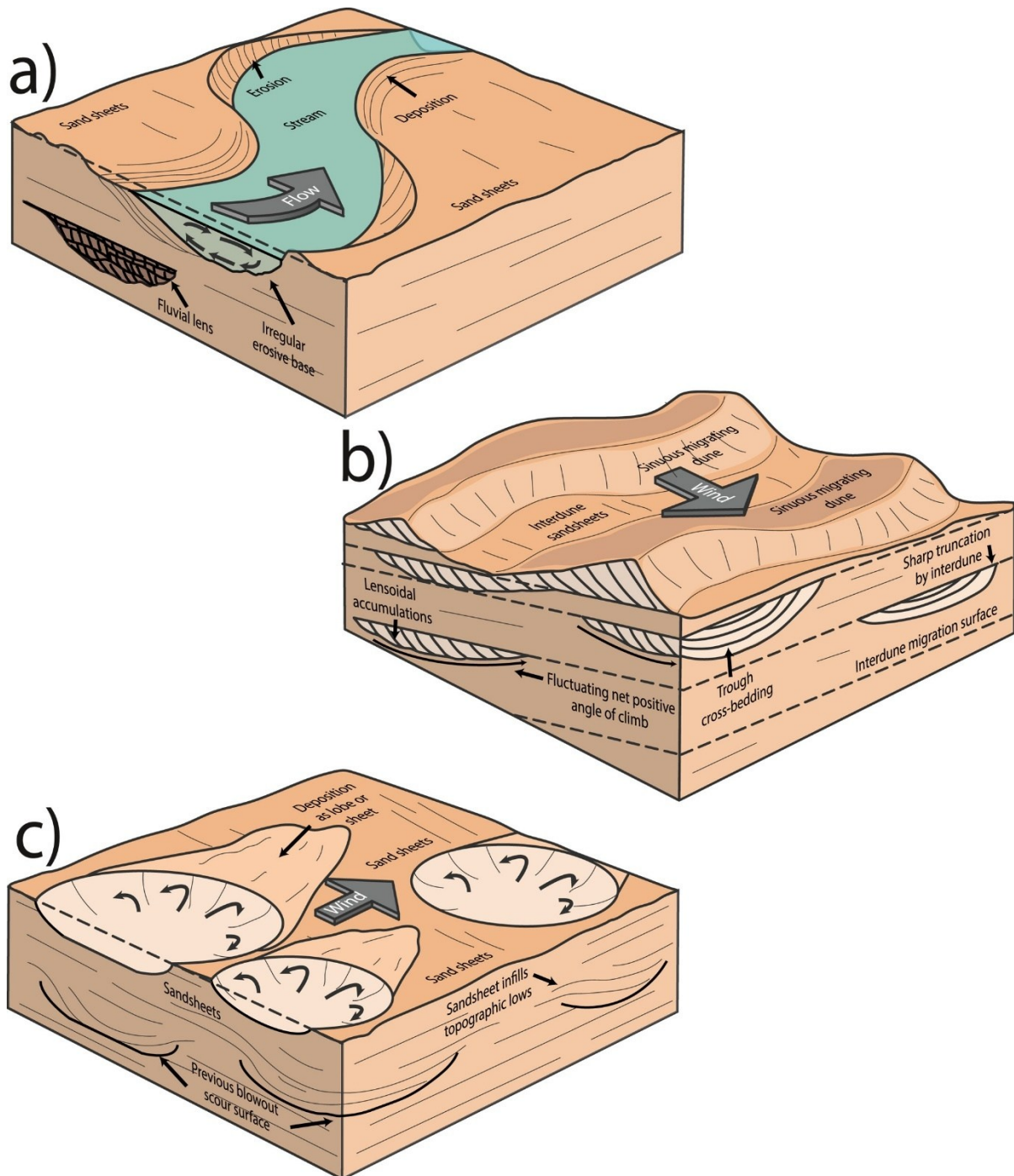


Figure 2.19: Schematic showing scenarios for scour-and-fill structure formation. (a) Scouring through aqueous processes. (b) Scouring through bedform migration. (c) Scouring through localized wind deflation.

2.7 Scour-and-fill structures as evidence for intensification of wind regime

Although most of the literature on the morphology of blowouts focuses on biotic and anthropomorphic processes driving blowout initiation, one of the key abiotic precursor processes for blowout formation are prolonged or episodic intense and undersaturated winds (Hesp, 2002). In regions with little to no vegetation, Hesp and Hyde (1996) observed that 1-3 m wide to 0.5-1 m deep saucer blowouts were initiated during a 10 hour long gale. Subsequent field studies have also identified high velocity winds as integral to blowout morphology and development in sparsely vegetated regions (Abhar et al., 2015; Levin, 2011; Luo et al., 2019a; Luo et al., 2019b). Modelling which used parameters for a stabilised surface also found high wind speeds to be an effective exogeneous disturbance capable of forming blowouts (Barchyn and Hugenholtz, 2013). Plausible parameters which could promote surface stabilization could be grain armoring (e.g. Blake et al., 2013; Weitz et al., 2018), surface cementation (Nickling, 1984), or fluctuations in near-surface wetness (Wiggs et al., 2004). Modelled storm events show that wind speed within a blowout after initiation will increase above the sediment carrying threshold and lead to erosion (Smyth et al., 2013). The extent of erosion strongly increases with higher wind velocities, showing a positive correlation between wind speed and erosional intensity (Smyth et al., 2019). I suggest that the observed scour-and-fill structures may have resulted from localized surface deflation due to the interplay between high-velocity, low sediment-saturation wind activity with a stabilized or partially stabilized surface. While the exact timescale for such processes on Mars remains uncertain, analogous blowout development on Earth typically occurs over decades (Abhar et al., 2015; Hugenholtz and Wolfe, 2006; Kayhko, 2007; Luo et al., 2019b) and even centuries (Luo et al., 2019a) suggesting a potentially similar timeframe under Martian conditions.

The stratigraphy prior to the scour-and-fill structures in the Mirador formation is dominated by aeolian deposition, characterized by sand sheets and aeolian dunes (Edgar et al., 2024; Gupta et al., 2022; Rapin et al., 2021; Roberts et al., 2024). The transition to an erosive regime marked by significant wind deflation suggests a shift in environmental conditions, likely driven by an increase in wind velocity and/or a decrease in wind sediment-saturation. These scour-and-fill structures have not been observed in the lower stratigraphy of the Mirador formation and are restricted to a single stratigraphic interval (-3829 to -3841 m), after which aeolian deposition in a sand sheet environment resumes again (Section 4; Roberts et al., 2023). This isolated stratigraphic position suggests that the conditions which formed the scours were constrained to a specific time interval, an interpretation that naturally leads to the inference that there was a period of sustained, higher-intensity wind conditions in Gale crater which promoted scour formation.

2.8 Conclusions

This chapter documents decametre-wide, concave-up scour-and-fill structures in the Chenapau member of the Mirador formation - the basal stratigraphic interval of the orbitally defined layered sulfate-bearing unit - exposed in the lower foothills of Aeolis Mons, Gale crater. Detailed analysis of their geometries, infill architecture, and associated facies reveals two distinct types of scour-and-fill structures, reflecting separate formative processes. The stratigraphy of the Chenapau member is interpreted as aeolian, characterized by planar, laterally extensive laminations with a pin-stripe pattern indicative of wind-ripple migration. The tabular nature, planar stratification, and lateral continuity of these deposits suggest deposition in a wind-transported sand sheet, where near-surface water inhibited dune formation. The two distinct types of scour-and-fill structures can be grouped into two classes: (A) broad, simple scours with a light-toned draping infill which reach up to ~60 m in width; and (B) small scours with lateral pinching and a dark-toned appearing infill. A singular scour that represents scour class B was identified in our research area. I interpret this class B scour as a fluvial scour with preserved channel wings and an infill comprised of cm-scale, tabular cross-stratification, which is characteristic of fluvial dunes. This scour represents a distinct episode of fluvial activity, unrelated to the processes responsible for Class A. In contrast, the geometry of Class A scour-and-fill structures - including broad and symmetrical cross-sectional concave profiles and flat central portions - and their overfilled, light-toned infill, continuous with the surrounding sand-sheet stratigraphy, suggests they are aeolian in origin. More specifically, the geometry, architecture, and infill facies are consistent with blowout structures. These blowouts likely formed due to localised deflation of a stabilized surface by increased wind velocity and/or decreased wind sediment saturation. Because these Class A scour-and-fill structures occur only within a single stratigraphic interval and share comparable elevations, I propose that their formation reflects a distinct, temporally confined shift in environment from deposition of the surrounding aeolian sand sheets. Collectively, the Class A Scours provide evidence for a regional event of higher wind velocity during the formation of the layered sulfate-bearing unit.

Chapter 3: An Aeolian Depositional Sequence Shaped by Near-Surface Water at the Base of the Layered Sulfate Unit, Gale Crater, Mars

3.1 Introduction

Across Mars, younger sulfate-bearing sedimentary successions are often found to overlie older, clay-bearing strata (e.g. Ehlmann and Edwards, 2014), leading to a hypothesis that a global change in climate associated with the aridification of Mars, recorded this broad shift in mineralogy (e.g. Bibring et al., 2006; Ehlmann and Edwards, 2014). Sedimentary successions are the only physical record available to understanding how Mars evolved from a once “warm and wet” paleoclimate, potentially suitable for supporting life, to the “cold and dry” climate present on Mars today (Fairén, 2010; Grotzinger et al., 2014; Hurowitz et al., 2017; Malin and Edgett, 2000). Aeolis Mons – the central mound within Gale crater - is an exceptional example of a stratigraphic succession recording this global climate transition (Fraeman et al., 2016; Milliken et al., 2014; Sheppard et al., 2021b). Gale crater was selected as the landing site precisely because the central mound preserves a thick, accessible record of this transition (Golombek et al., 2012). One goal of the Mars Science Laboratory (MSL) mission is to use *Curiosity* to test whether this shift in mineralogy is linked to an overall aridification of the region and whether habitable environments could persist through periods of climatic change (e.g. Golombek et al., 2012; Milliken et al., 2010; Vasavada, 2022). So far, rover observations of lacustrine and fluvial deposits in the clay-bearing unit, and aeolian deposits with interbedded fluvial/lacustrine lenses in a mixed clay–sulfate unit, support this overall regional “drying out” (Caravaca et al., 2025; Caravaca et al., 2022a; Caravaca et al., 2022b; Edgar et al., 2020; Grotzinger et al., 2014; Gupta et al., 2023a; Rivera-Hernández et al., 2019; Stack et al., 2019).

The orbitally defined Layered Sulfate unit (LSu) which forms part of this succession is composed of distinct, planar-parallel, laterally extensive layers that are visible from orbit (e.g. Anderson and Bell III, 2010; Fraeman et al., 2016), analogous to other thick successions of layered sulfate terrain across Mars (Gendrin et al., 2005). Multiple formational mechanisms – from depositional processes to diagenesis post-dating lithification – have been suggested to explain both the mineralogy and appearance of these layered sulfates (e.g. Ehlmann and Edwards, 2014; Gendrin et al., 2005). In 2022, *Curiosity* explored the basal section of the LSu in a canyon where a ~100 m thick succession is exposed in local buttes. This stratigraphic succession represents an ideal location to study the depositional environments at the base of the LSu due to the excellent outcrop exposures of the sedimentary architecture, and for the lateral continuity of the layers across the rover’s traverse, allowing for on-ground facies characterisation.

The aim of this work is to reconstruct the depositional environments and processes that created the sedimentary succession of the basal section of the layered sulfate-bearing unit. The main objectives are to: (1) describe and characterize the sedimentary structures, facies, and architecture recorded in the basal section of the layered sulfate unit; (2) reconstruct the depositional processes and environments for each member; (3) evaluate the evolution of these depositional environments over the formation of this succession, and (4) discuss the implications for the formation of the layered sulfate-bearing unit at Gale crater and their broader, regional significance.

3.2 Geologic Context

Gale crater (Figure 2.1a) is a partially infilled impact crater approximately ~155 km in diameter, located within the heavily cratered southern highlands near the dichotomy boundary with the comparatively smoother northern lowlands (Tanaka et al., 2014). Crater counting of its ejecta blanket places its formation at around 3.7-3.8 billion years (Ga) during the Noachian-Hesperian transition period (e.g. Le Deit et al., 2013; Thomson et al., 2011). Following its formation, Gale was progressively infilled by sedimentary deposits over a 200 to 500 million year (Myr) period (e.g. Grant et al., 2014; Thomson et al., 2011), punctuated by several episodes of aeolian erosion (Dromart et al., 2021). The infilling material in Gale crater was later exhumed by predominantly aeolian processes (Anderson and Day, 2017; Day and Kocurek, 2016; Grotzinger et al., 2015; Kite et al., 2013b; Malin and Edgett, 2000) leaving behind the ~5.5 km tall central mound, Aeolis Mons. This mound, informally known as “Mount Sharp,” is comprised of distinctive layered stratigraphy (Anderson and Bell III, 2010; Malin and Edgett, 2000) (Figure 2.1a). Reflectance spectral data from orbit indicate a transition up-section from clay-bearing to layered sulfate-bearing strata, which are themselves unconformably overlain by anhydrous mineral-bearing strata (Figure 2.1b; Fraeman et al., 2016; Milliken et al., 2014). This clay-to-sulfate transition is found to be a spatially variable and inconsistent gradational boundary, with discontinuous inclusions of clay-bearing strata amongst the layered sulfate-bearing strata from further orbital studies (Sheppard et al., 2021b).

By August 2025, *Curiosity* has traversed over 850 m of vertical stratigraphy since its landing in August 2012 (Figure 2). Three major stratigraphic groups have been documented at the time of writing: the Bradbury, Mount Sharp, and Siccra Point groups. The Bradbury group consists of fluvial sandstones and conglomerates outcropping across the crater floor and the lower slopes of the mound (Aeolis Palus) (Rice et al., 2017; Stack et al., 2016; Vasavada, 2022; Williams et al., 2013). The Bradbury group is inferred to interfinger with strata of the Mount Sharp group (Grotzinger et al., 2015). The Mount Sharp group consists of stratigraphy which bears hydrated minerals identified from orbit prior to landing (e.g. Milliken et al., 2010). These groups are unconformably overlain by the Siccra Point group. The Stimson formation is the only unit assigned to the Siccra Point group so far (Banham et al., 2022; Watkins et al., 2022), which is composed of large-scale aeolian cross-bedded units indicating an arid desert environment (Banham et al., 2018), although minor soft sediment deformation structures show that

near-surface fluid was locally present before lithification (Banham et al., 2024b).

The Mount Sharp group is comprised of three formations (Figure 2.2). The Murray formation is comprised of clay-bearing strata with lacustrine mudstones and subordinate aeolian and fluvial sandstones (Edgar et al., 2020; Grotzinger et al., 2014; Rivera-Hernández et al., 2019; Stack et al., 2019). This is conformably overlain by the Carolyn Shoemaker formation which contains interbedded mudstones and cross-bedded sandstones interpreted as an intermittent fluvial and lacustrine environment (Caravaca et al., 2022b; Cardenas et al., 2022; Fedo et al., 2022; Gwizd et al., 2024). The Carolyn Shoemaker formation overlies the Mirador formation, possibly unconformably (e.g. Meyer et al., 2025), and corresponds to where the orbitally identified layered sulfate-bearing strata is observed from orbit (Fraeman et al., 2016).

The Mirador formation is currently subdivided into 6 members (Figure 2.2): the Dunnideer, Port Logan, Contigo, Catrimani, Amapari, and Chenapau members. Metre-scale cross-sets occur at the base of the Mirador formation (Rapin et al., 2021), specifically in the Dunnideer, Port Logan, and Contigo members with varying degrees of diagenetic alteration (Edgar et al., 2024; Meyer et al., 2025; Rapin et al., 2023b). The Contigo member also contains isolated, resistant lenses that record intermittent fluvial and lacustrine deposition (Caravaca et al., 2022a; Gupta et al., 2023b). The Catrimani member exhibits planar, laterally extensive laminae interpreted to be from the migration of wind ripples in aeolian sand sheets (Edgar et al., 2024; Roberts et al., 2023). Sharply overlying the Catrimani member is the thin (~0.5 m thick) Amapari member – originally identified from orbit and called the marker band for its distinctive dark-toned appearance, and laterally extensive trace (Weitz et al., 2022) – containing symmetric ripples interpreted to be formed in a shallow lake (Gupta et al., 2023a; Mondro et al., 2025a; Mondro et al., 2025b; Thompson et al., 2024; Weitz et al., 2023). The Amapari member is then overlain by the Chenapau member which consists of planar strata and is truncated by the large-scale scours interpreted to be from aeolian scouring of sandsheet deposits (Roberts et al., 2025). Mg-sulfate was first identified by CheMin within the upper-most section of the Contigo member (Chipera et al., 2023), marking it the base of the orbitally defined Layered-Sulfate unit (LSu). The LSu therefore coincides with the upper-most section of the Contigo member, along with the Catrimani, Amapari, and Chenapau members which I discuss in depth in this manuscript.

3.3 Methodology and data sets

3.3.1 Instruments and data products

Architecture and unit relationships of the stratigraphy exposed in Marker Band valley was investigated primarily using far-field images (>50 m) taken by Mastcam, Navcam, and ChemCam instruments aboard the *Curiosity* rover. When stratigraphic units extended outside of the range of the rover's viewshed, stratigraphic unit correlation was investigated using orbital imaging data products. Detailed analysis of the sedimentary facies was performed using near-field images (<50 m) taken by

the Mastcam, ChemCam, and MAHLI instruments.

3.3.1.1 Rover instruments and data products

The two Mast Cameras (Mastcams - left and right; Bell et al., 2017; Malin et al., 2017) are located ~1.9 m above ground level. High resolution RGB colour images are provided by the 0.075 mrad/pixel sampling of the M100 camera (~3.7 mm/pixel at 50 m viewing distance). Stereo photomosaics are possible due to the ~24.2 cm separation between the M34 and M100 cameras, enabling spatial sampling of 0.22 mrad/pixel (~10.9 mm/pixel at 50 m viewing distance). A white colour balance (see Appendix III.3) was performed using *Adobe Photoshop* to ensure colour consistency between images, to highlight subtle contrasts that record changes in surface properties, and to provide Earth-like lighting conditions for better representation of colours for interpretation and comparisons with terrestrial analogues.

The Navigation cameras (Navcams; Maki, 2018; Maki et al., 2012) provide 360°, monochromatic (greyscale) context stereo mosaics at every rover waypoint with a resolution of 0.82 mrad/pixel (41.0 mm/pixel at 50 m viewing distance). With an interocular length of ~42.4 cm between the two navigation cameras, stereo 3D-mesh products were derived (Alexander et al., 2006) typically accurate to a distance of ~30-50 m from the rover.

The Chemistry and Camera instrument suite (ChemCam; Maurice et al., 2016) has the capability to take long distance image mosaics with its Remote Micro-Imager (LD RMIs; Le Mouelic et al., 2015; Le Mouelic et al., 2019). At 0.0205 mrad/pixel (1.0 mm/pixel at 50 m viewing distance), LD RMIs were sometimes used to supplement Mastcam data where instrument pointing safety restrictions allowed. Panchromatic sharpening of high-resolution LD RMIs using lower-resolution colour Mastcam photomosaics was occasionally performed in *ArcGIS Pro 3.1.1* to enhance colour detail at certain outcrops outside of the rover's traverse.

The Mars Hands Lens Imager (MAHLI) is attached to *Curiosity's* arm turret and can be maneuvered to take close-up colour images of outcrops. Standard procedures involves a sequence of images taken at nominal distances of 1 cm (spatial sampling of 16 to 21 μm per pixel), 5 cm (spatial sampling of 31 μm per pixel), and 25 cm (spatial sampling of 100 μm per pixel) (Edgett et al., 2015; Edgett et al., 2012; Yingst et al., 2016). "Dog's eye" observations orient MAHLI orthogonal to the sedimentary structure within the outcrop. MAHLI has the ability to resolve grain sizes of coarse silt, ~62.5 μm , or larger (Edgett et al., 2015; Yingst et al., 2016).

3.3.1.2 Orbital Data Products

The rover-derived data products were supplemented with ~25 cm/pixel orbital images from the High Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007). For 3D data, a ~1 m/pixel Digital Terrain Model (DTM) generated from HiRISE stereo pairs and registered to the Mars Orbiter Laser Altimeter data products (Calef and Parker, 2016; Smith et al., 2001) was used.

3.3.2 Data Collection and Imaging Strategy

3.3.2.1 Far-field imaging strategy

The rover entered Marker Band valley (Figure 3.1), the study area of this manuscript, on Sol 3575 (number of Martian days since *Curiosity* first landed in Gale crater). Marker Band valley is bracketed by the buttes informally named Orinoco, Chenapau, Bela Vista, Deepdale, and Bolivar (Figure 3.1) all of which contain portions of the ~100 m thick sedimentary succession described in this manuscript. As the rover traversed along an optimal route to navigate through the terrain - known as the Mount Sharp Ascent Route (MSAR), photomosaics of these buttes were planned at multiple angles to maximise coverage along the rover's closest approach when permitted by time, data volume, and energy constraints (e.g. Vasavada, 2022).

3.3.2.2 Geologic map creation for strategic planning

Distinct alternating 'dark-toned' and 'light-toned' stratigraphic packages - tonal differences that were also visible in HiRISE mosaics (Figure 3.1c) - were identified on each butte. These stratigraphic packages have different sedimentary textures, weathering patterns, and elevations visible from surface and orbit, which allowed for the subdivision of this stratigraphic succession into 11 distinct units (as described in section 4). To spatially contextualise these units, I created shapefiles of unit outcrops through comparisons of Mastcam mosaics with 3D local scene views in *ArcGIS Pro 3.1.1*, employing the same coordinates, elevation, heading, roll, pitch and extent parameters as the rover's mast camera at each location (Table II.6 in Appendix II summarises elevations for all stratigraphic units across the major buttes and along the rover traverse). I could then correlate these units identified in Mastcam mosaics to distinct tones, surficial textures, apparent induration, and elevations of rocks in HiRISE mosaics. Only a small amount of vertical elevation changes were identified for each unit between nearby buttes consistent with the gentle dips of 2-4° for bedding in the Mount Sharp group (Stein et al., 2020; Turner and Lewis, 2023). A geological map was created for this study area and these stratigraphic units were projected across the planned traverse (Figure 3.1d) allowing the rover to document the lateral equivalent of each unit for workspace imaging and contact science.

3.3.2.3 Rover workspace imaging and contact science

The rover traversed across each of these stratigraphic units from Sols 3575 to 3919 (Figure 3.1c). The geologic map aided contact science and outcrop selection along the MSAR when engineering constraints allowed (e.g. Rankin et al., 2020). Representative outcrops were identified in Navcam mosaics and targeted with Mastcam for more detailed facies analysis when planning constraints allowed. MAHLI image coverage had additional planning constraints associated with deployment of the arm, a process which typically takes 2 Sols to complete (Yingst et al., 2016). In total, over 150 workspace images were used in identifying the sedimentary textures and lithofacies of each unit as described in section 4.

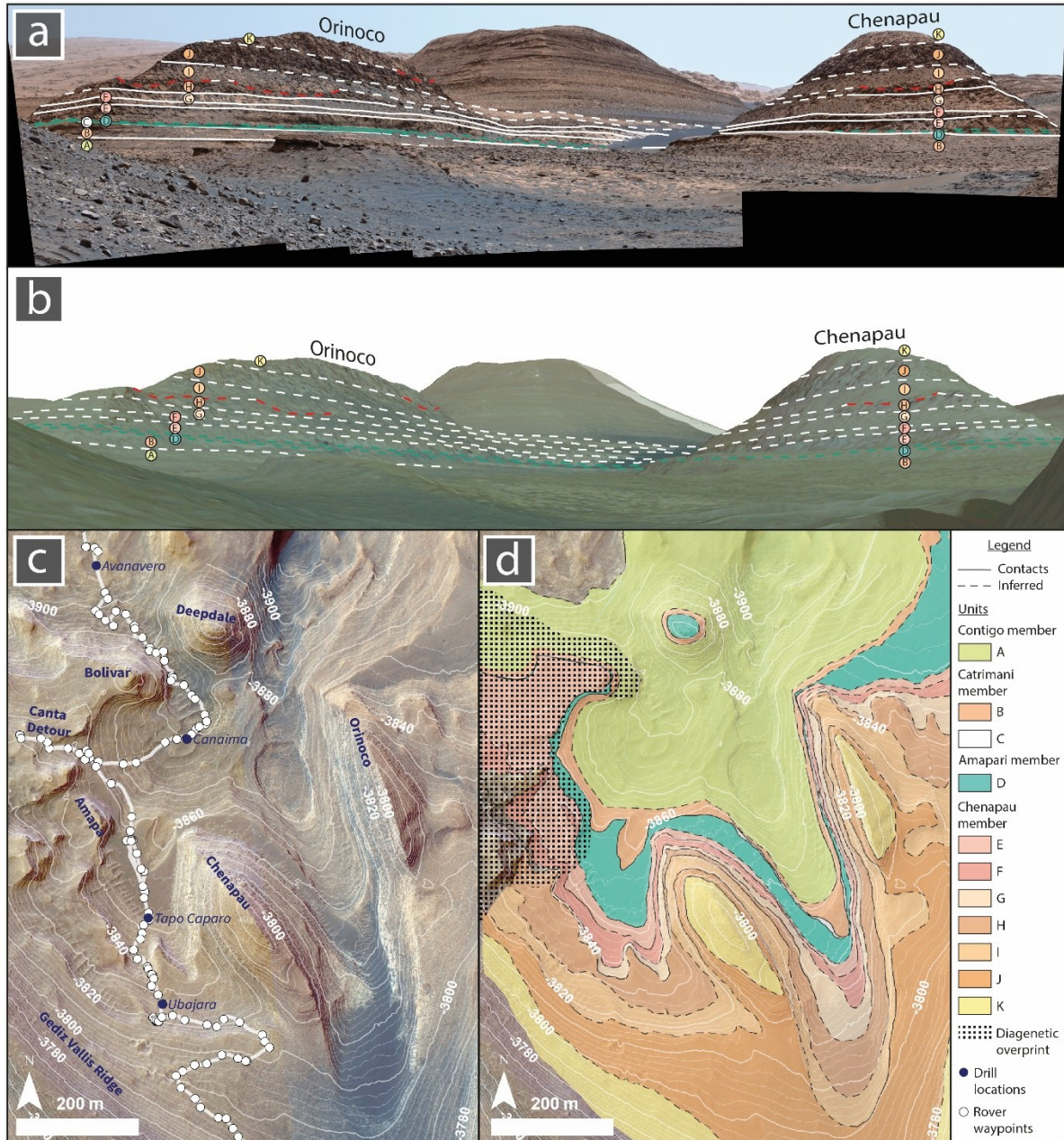


Figure 3.1: Geologic context used in planning. (a) Overview mosaic showing mapped geologic units in the Orinoco and Chenapau buttes in the Marker Band valley study area used in planning. Contact boundaries (white). The Amapari member (unit D) is in blue. Erosional scours are in red. Inferred boundaries are dashed. (Mastcam mosaic MR_mcam102554, acquired on Sol 3562, at Sol 3650 end-of-drive location). (b) Annotated arcgis pro scene view of (a) showing the High Resolution Imaging Science Experiment (HiRISE) mosaic (~25 cm per pixel; Calef and Parker, 2016) draped across the the HiRISE derived digital terrain model (DTM) (~1 m per pixel resolution; Calef and Parker, 2016; Smith et al., 2001). (c) Overview HiRISE map of Marker Band valley region of the rover's traverse (from sols 3504 to 3910). White line shows Curiosity's traverse. Black-rimmed circles show the key imaging locations. Blue-filled circles show drilling locations. (d) Geologic map of the Marker Band valley region used in planning. Image credits: NASA/JPL-Caltech/MSSS.

3.3.3 Data analysis

3.3.3.1 Mastcam image analysis

Mastcam data products were processed to optimise both visibility and accuracy of the sedimentary textures and architecture for analysis following Banham et al. (2018). Analysis of the stratigraphy was predominantly performed using stereo pair images from M34 and M100 (Table II.2, Appendix II). Interpreted data products used M100 mosaics as the base image. As the range to outcrops varied relative to the rover (Figure I.1, Appendix I), different approaches were required to assess sedimentary textures at different scales. For “farfield” outcrops - predominantly coverage of the stratigraphic cross-sections exposed in the local buttes - spatial sampling ranged from ~17 to 0.5 mm/pixel and were only used to examine bulk stratigraphic architecture, sedimentary architectures (if visible at that scale), unit contacts, and location and extent of unit outcrops. For “nearfield” outcrops – coverage of outcrops near to or within the rover workspace - spatial sampling ranged from 0.5 to 0.1 mm per pixel and were used for more detailed facies analysis, and sedimentary architecture. Mosaics were interpreted using vector graphic software to highlight facies, sedimentary architecture, and relationships between depositional surfaces.

Quantitative information on sedimentary architecture and facies was derived from the 3D Navcam mesh (Alexander et al., 2006) when outcrops were within ~50 m. At distances further than 50 m, quantitative information was provided using *ArcGIS Pro 3.1.0* onto the HiRISE orthomosaic and used as a base to correlate features of interest visible in both the HiRISE and Mastcam mosaics. The GIS projected viewsheds (see Appendix III.2) using the HiRISE DTM based on the rover co-ordinates and imaging specifications were also aided mapping geologic units and features of interest. To account for foreshortening when measuring apparent dips, I only measured the parts of the image mosaic oblique to outcrops.

3.3.3.2 MAHLI image analysis

MAHLI images helped resolve fine-scale sedimentary structures in the stratigraphy for a more detailed facies characterization (a list of MAHLI images used in this chapter is listed in Table II.4, Appendix II). Measurements of laminae thickness could be performed on MAHLI images orthogonal to bedding as described by Edgar et al. (2020). Grayscale pixel intensity was assessed as a proxy for laminae thickness as described by Gwizd et al. (2022) using the open-source image processing program, ImageJ (Schneider et al., 2012). Python code was developed using the scipy library (Virtanen et al., 2020) using the “find_peaks” function on the grayscale pixel intensity with the difference between peaks used as a proxy for laminae thickness as described in S1 in Grotzinger et al. (2015). The arithmetic mean, median, minimum, maximum, and standard deviation values for each measured section were then calculated.

3.4 Stratigraphy and Sedimentology

The Contigo, Catrimani, Amapari, and the Chenapau members are exposed as a ~100 m thick sedimentary succession across the seven buttes that define Marker Band valley (Figure 3.2). Within this succession, eleven distinct geologic units, designated units A through K from oldest to youngest, were identified based on distinct bedrock textures, observable contacts, sedimentary architecture, and stratigraphic distribution (summarised in Table 3.1). Here, I describe the stratigraphic architecture and facies of units A through K and provide interpretations of their depositional processes in stratigraphic order.

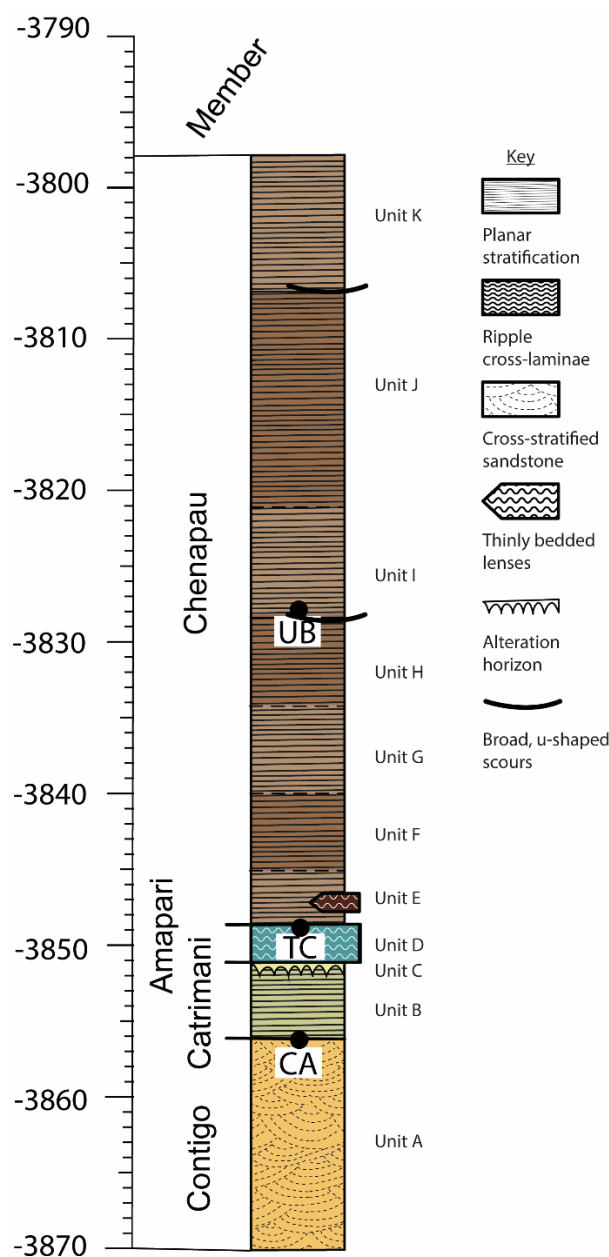


Figure 3.2: The stratigraphic column of the Contigo (yellow), Catrimani (green), Amapari (blue), and Chenapau (brown) members of the Mirador formation and units A through K as discussed in this chapter. Note: the light- and dark-toned planar bedded lithofacies in the Chenapau member are represented by light and dark brown shading respectively.

Unit	Thickness (m)	Dominant Stratification	Architectural Elements	Unit Contact	Lithofacies	Interpretation
A	~40	Low angle cross-bedding	Vertically stacked, laterally continuous tabular beds (2–6 m thick) of low-angle cross-stratification, dipping north, abutting sharp planar bounding surfaces.	Irregular contact with underlying clay-sulfate transition; ~1 m paleo-topographic relief.	Interbedded resistant dark-toned laminae and cm-scale light-toned sets at bed base; top composed of homogeneous, laterally extensive planar laminae.	Lateral accretion of simple migrating aeolian dunes. The irregular lower contact likely marks a regionally significant aeolian supersurface above the clay-sulfate transition.
B	~5	Planar, light-toned stratification	Laterally extensive tabular bed; consistent thickness over ~350 m.	Sharp, planar boundary with unit A.	Uniform planar laminae; pervasive nodules (~3–4 mm).	Continuous deposition in a dry aeolian sandsheet environment with limited sand supply and near-surface water.
C	~0.3	Structureless, mottled light-toned	Mottled outcrop with cycloidal lower contact pattern; infills of V-shaped wedges.	Irregular lower contact forming a cycloidal pattern against unit B.	Cycloidal base pattern; south-dipping fractures; nodular mottled uppermost ~0.05 m	Contraction-expansion processes potentially thermal contraction on a regional scale above a supersurface with surface alteration, possibly linked to groundwater fluctuations.
D	~1	Lower dark-toned resistant; upper light-toned recessive	Tabular ledge/platform forming unit; lateral continuity up to 250 m+.	Planar lower contact with underlying unit C	Lower: symmetric ripple forms (~4.5 cm wavelength) in chevron stacks; Upper: thick, continuous laminations thinning upward.	Lacustrine deposition in a deepening lake.
E	~3.6	Planar, even-in-thickness laminae	Scarp-forming tabular bed; sharp, planar basal boundary with subtle overhang; lateral continuity ~125 m. Embedded dark-toned lens with cm-scale cross-bedding and lateral pinching.	Sharp basal, planar contact marked by a resistant band.	Uniform planar bedding; laminae overprinted by diagenesis.	Subaerial deposition – likely as an aeolian sandsheet - following a transition from lacustrine conditions. Embedded dark-toned lens likely a fluvial lens.

F	~4.2	Planar stratification with corrugated textures.	Tabular bed; consistent thickness across ~350 m; corrugated textures present.	Sharp conformable basal contact with unit E.	Interbedded light/dark planar strata; corrugated horizons; preserved ripple forms with lee/stoss sides; mm-scale nodules.	Deposition of supercritically climbing ripples that can be attributed to either aeolian or damp conditions.
G	~2.7	Planar, light-toned laminations	Slope-forming, recessive unit with homogenous appearance and rounded weathering profile.	Gradational contact from underlying unit F; absence of wavy bedding.	Planar, low-angle to horizontal laminations; resistant “fins” or protrusions creating corrugated textures; homogeneous light tone; mm-scale nodules dispersed.	Continuous deposition in a dry aeolian sandsheet environment with limited sand supply and near-surface water.
H	~5.3	Cm-scale sets of recessive-resistant packages of planar strata.	Scarp-forming unit of recessive sets interbedded with thin resistant horizons; laterally continuous (~250 m extent).	Gradational contact with unit G; marked by introduction of indurated beds.	Planar, parallel laminae in sets (1–12 cm thick) with “pinstripe” light/dark pattern; corrugated resistant laminae forming fins (~4–5 cm spacing); coin-shaped fragments near top. Occasional truncation surfaces. Wavy and irregularly undulating laminae. Occasional frozen ripples.	Intermittent damp aeolian sandsheet deposition with candidate adhesion structures. Indurated laminae suggest preferential cementation, likely formed by later diagenetic processes.
I	~4.9	Uniform, light-toned, even-in-thickness laminae	Light-toned, smooth, rounded blocks; often appears massive due to diagenetic overprint.	Drapes over unit H with associated scours; nature of most contacts uncertain.	Evenly thick planar laminae (~0.76 mm) with light/dark pinstripe pattern; mm-scale nodules; occasional small scours; absence of flaky resistant laminations.	Continuous deposition in a dry aeolian sandsheet environment with limited sand supply and near-surface water.
J	~7.3	Cm-scale sets of recessive-resistant packages of planar strata	Dark-toned, interval comprised of resistant and recessive interbedded beds; laterally continuous with minor truncation by large scours.	Gradational contact with unit I; marked by introduction of indurated beds.	Dark-toned, indurated laminae interbedded light-toned pinstripe laminae into 1–12 cm sets; dark-toned, corrugated, resistant laminae form parallel fins/ridges ~4–5 cm apart. Low-angle truncation.	Intermittent damp aeolian sandsheet deposition with candidate adhesion structures. Indurated laminae suggests preferential cementation, likely formed by later diagenetic processes.

K	~12.9	Uniform, light-toned, even-in-thickness laminae	Slope-forming; light-toned, smooth, rounded blocks with abundant mm-scale concretions.	Overlies unit I with sharp contacts where visible.	Uniform, laterally extensive planar laminae; consistent “pinstripe” alternating light/dark pattern; mm-scale nodules.	Continuous deposition in a dry aeolian sandsheet environment with limited sand supply and near-surface water.
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Table 3.1: Sedimentary facies for units A through K

Member	Unit	Apparent laminae thickness (mm)	s.d (mm)	Count	Resolution	Image
Contigo	A	~0.7	0.3	219	0.099 mm/pixel	Sol3596_Marshall_Falls
Canaima	B	~1*	0.3	92	0.3 mm/pixel	Mcam102798
	C	N/A	N/A	N/A	N/A	Mamupi
Amapari	D	4 to 10	See Mondro et al (2025)			
Chenapau	E	~1**	N/A	N/A		Mcam103681
	F	~1**	N/A	N/A		Mcam103808
	G	~1*	0.3	34	0.5 mm/pixel	Mcam103899
	H	~3*	1.35	25	0.6 mm/pixel	Sol03807_Camopi_mcam_103951
	H/I contact	~1*	0.4	52	0.3 mm/pixel	Sol03819_ubajara_ext_stereo
	I	~0.76	0.058	180	0.1 mm/pixel	
	J	~1.5*	0.8	58	0.3 mm/pixel	Mcam104420, Zarelia
	K	<1 mm**	N/A	N/A		Mcam104608

Table 3.2: Measurements of Apparent Average Laminae Thickness for units A through K. This table presents the average laminae thickness for each unit, as determined through various methods. unit I was measured manually using analogue techniques; comparison with automated methods yielded identical average values but an increased standard deviation of 0.34 (See Figures I.3 and I.4, Appendix I).

* Measurements derived from Mastcam images, potentially affecting accuracy.

** units with poor outcrop exposure, lacking cross-sections of laminae.

3.4.1 Contigo Member

3.4.1.1 Unit A

Architectural elements

Unit A, approximately 40 m-thick, is characterized by repetitive, vertically stacked tabular beds of low-angle cross-bedding (Figure 3.3). Seven tabular beds (ranging in thickness from 2 to 6 m) were identified. These beds are laterally continuous, traceable for up to ~350 m along the western extent of the valley (Figure 3.3a), while maintaining consistent thicknesses (Figure 3.3b and c). Cross-bedding is evident in four of the beds, with strata sets dipping at apparent angles of ~4-5 degrees to the north. Each cross-bed was discordantly truncated by sharp, planar, laterally extensive bounding surfaces. While strata at the top of each cross-bed appeared light-toned and recessive, horizons of thin, dark-toned, erosion-resistant strata separate the light-toned strata sets higher up in the succession as they asymptotically tapered against the underlying bounding surface (Figure 3.3d and e). In northeast-facing sections, cross-bedding could not be identified (Figure 3.3e).

At similar elevations across the rover traverse (Figure 3.4), distinct planar bedding was identified separated by cm-thick, resistant, dark-toned bands (Figure 3.4a). Each bed was comprised of light-toned sets separated by thin, mm-scale, erosion-resistant, dark-toned laminae (Figure 3.4a and b). The light-toned sets consisted of parallel, laterally extensive laminae with apparent thicknesses of $\sim 0.7 \pm 0.3$ mm. Occasional weathered sections show thin, protruding, light-toned horizons between each laminae in a pinstripe pattern (Figure 3.4b and c). These light-toned laminae could be scratched by the Dust Removal Tool (DRT). The dark-toned strata truncated the sets of light-toned laminae and occasionally formed protruding fins (Figure 3.4b). In plan view, the resistant strata appeared dark grey, pitted, had a slight lustre in dust-free surfaces, and could not be scratched with the Dust Removal Tool (DRT) before MAHLI imaging. Given both the northward dip of the cross-bedding and how nearfield observations by the rover were made on a north-facing slope, the viewing angle was likely parallel to the strike of the cross-bedding, and thus perpendicular to the transport direction of these primary bedforms, thus explaining the absence of cross-stratification in nearfield images.

The lower contact of unit A was not observed in Marker Band valley due to poor exposure. However, an irregular surface was observed in lower Deepdale (Figure 3.5). This irregular surface, displaying ~1 m of paleotopographic variation, could be traced across outcrops for over 75 m visually distinct and recognisable in disparate outcrops. This surface sharply truncates underlying trough cross-stratification (Figure 3.5) and is conformably overlain by cross-stratified tabular beds with consistent set thicknesses and sharp planar bounding surfaces. As the sedimentary architecture was characteristic of unit A, and as these tabular beds occurred at similar elevations to exposures of unit A in Marker Band valley, I infer that this irregular surface marks the lower contact of unit A.

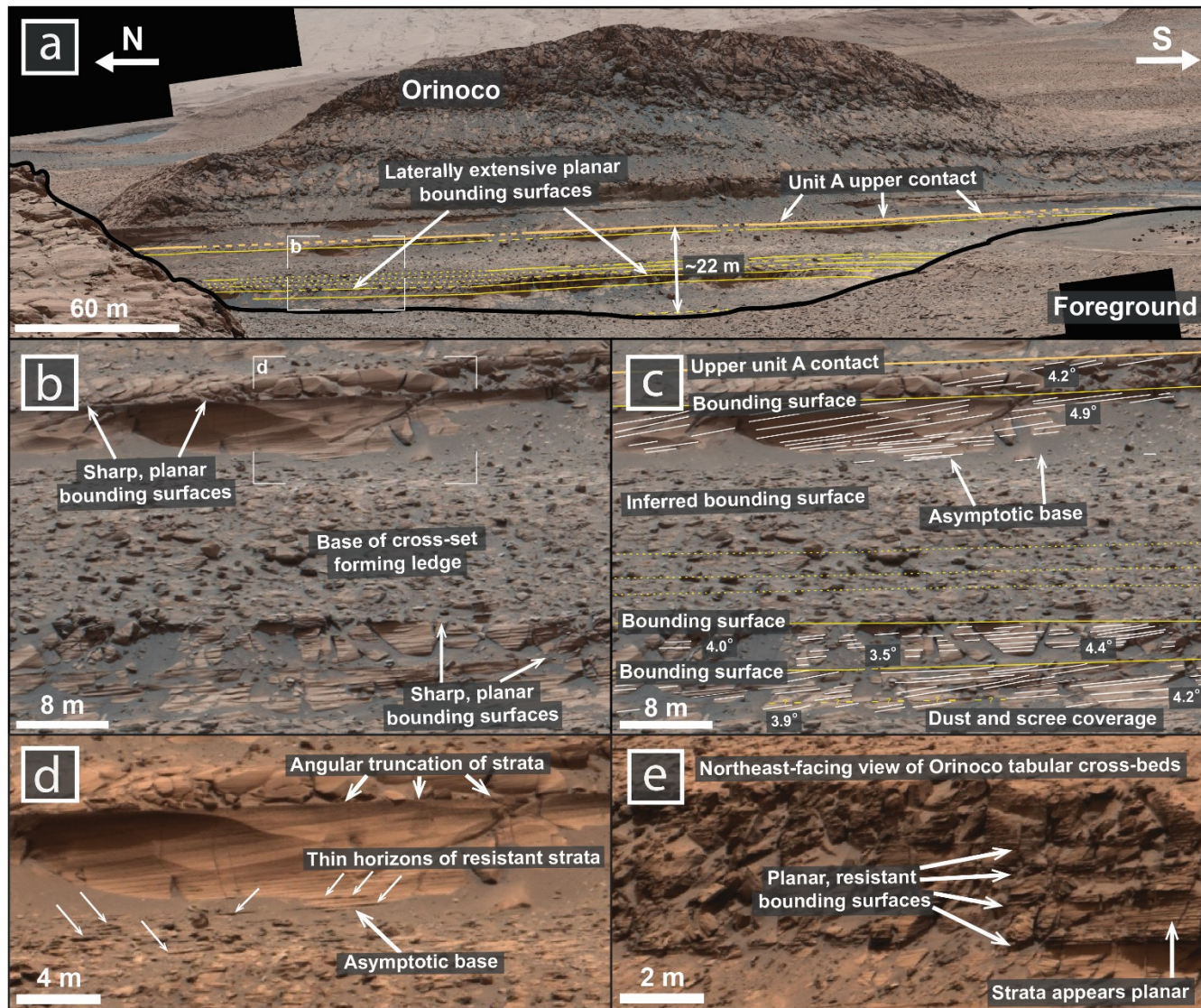


Figure 3.3: Orinoco butte mosaic showing unit A. (a) Mastcam mosaic of unit A illustrating unit A's upper contact (thick cream line) and bounding surfaces (yellow lines) where visible truncation of cross-strata (continuous), visible no truncation (dashed), and inferred (dotted) (mastcam mosaic MR_mcam103144 acquired on Sol 3662, at Sol 3658 end-of-drive location). (b) Detail of uninterpreted cross-bedded sets. (c) Interpretation of cross-bedded sets showing the upper contact, bounding surfaces, and cross-bedding (white). See Appendix III.4 for how the apparent dip was calculated. (d) Detail of cross-bedding showing angular truncation of strata, asymptotic bases of cross-sets, and thin horizons of resistant strata. (e) Northeast facing view of Orinoco tabular beds highlighting the absence of identifiable cross-bedding at different outcrop orientations and resistant strata concentrated at the planar bounding surfaces (mastcam mosaic MR_mcam103184 acquired on Sol 3667, at Sol 3665 end-of-drive location). Image credits: NASA/JPL-Caltech/MSSS.

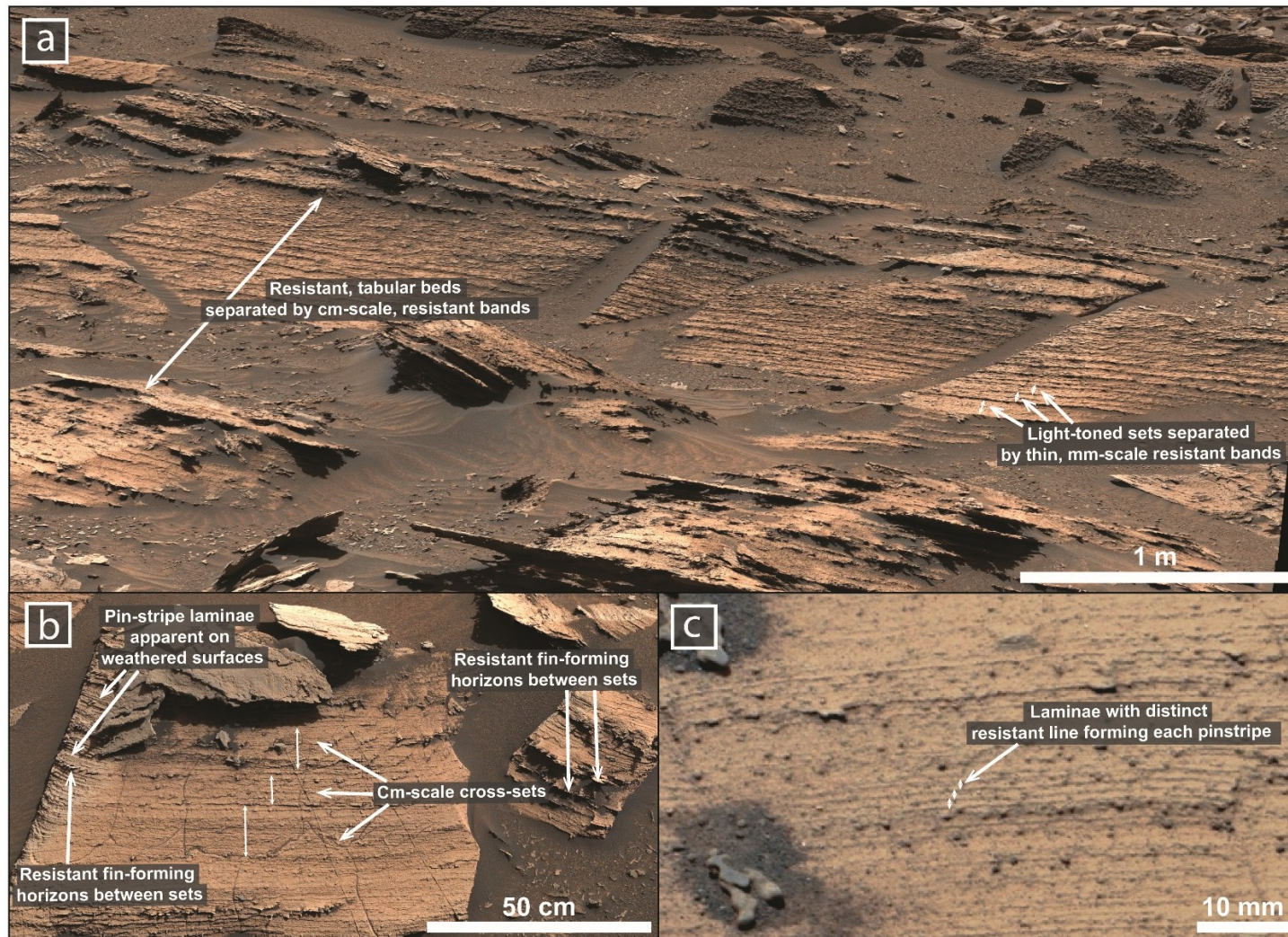


Figure 3.4: Examples of select facies characteristic of unit A as encountered across the traverse. (a) Mastcam mosaic of the tabular, m-scale bedding. Note the grouped, thick, dark-toned resistant bands that separate each bed. Beds are comprised of cm-scale sets separated by thinner mm-scale, resistant bands. (mastcam mosaic MR_mcam102699 acquired on Sol 3582, at Sol 3579 end-of-drive location). (b) Mastcam mosaic showing low-angle truncation of cm-scale sets. Note how resistant laminae between sets can form fins. (mastcam mosaic MR_mcam102748 acquired on Sol 3594, at Sol 3592 end-of-drive location). (c) Mastcam mosaic of the planar, laterally extensive, mm-scale laminae. Note how thin, resistant lines between each of the laminae form a pinstripe pattern. (mastcam mosaic MR_mcam102757 acquired on Sol 3596, at Sol 3594 end-of-drive location). Image credits: NASA/JPL-Caltech/MSSS.

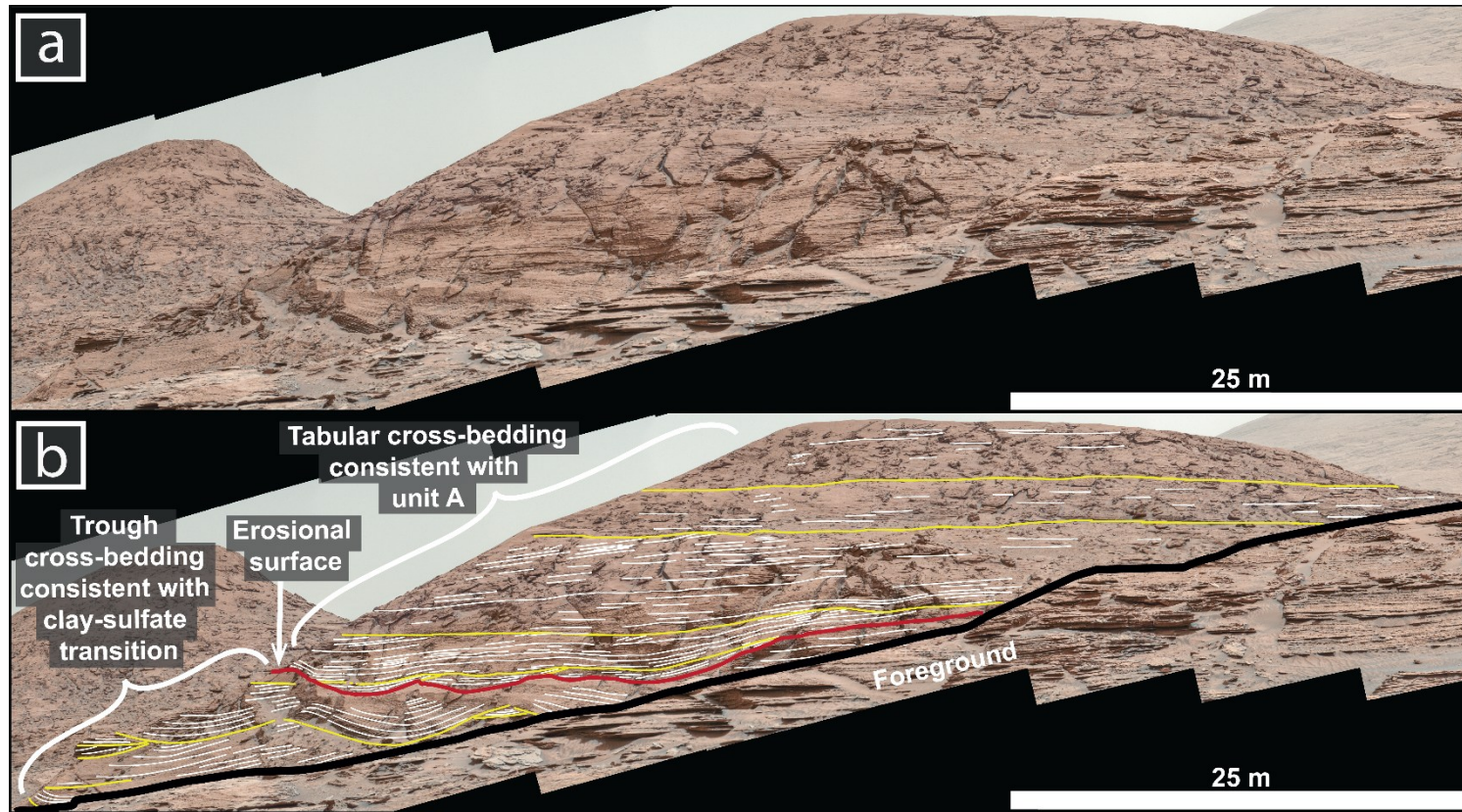


Figure 3.5: Mastcam mosaic of inferred boundary between the underlying Clay-Sulfate transition and the overlying unit A. (a) the uninterpreted mosaic of the butte north of Deepdale (mastcam mosaic MR_mcam102097 acquired on Sol 3485, at Sol 3483 end-of-drive location) and (b) the interpretation of the erosional surface that separates trough cross-bedding consistent with bedding expression in the clay-sulfate transition (Gupta et al., 2023; Edgar et al., 2024) and tabular cross-bedding consistent with the lithofacies observed in unit A. This erosional surface (red) expresses ~1 m of palaeo-topographic relief. Bounding surfaces (yellow lines) separate and truncate cross-strata (white lines). Note how overlying bounding surfaces and stratigraphy drape across the erosional surface. Image credits: NASA/JPL-Caltech/MSSS.

Interpretation

The geometry of the vertically stacked, north-dipping cross-beds suggest that the deposits originated from lateral accretion of simple, migrating aeolian dunes. As the beds appear planar and lack observable cross-bedding in south- and north-facing outcrops, the primary bedforms were likely simple, two dimensional, and migrating towards the north (e.g. figure 5 in Rubin and Hunter, 1982). Due to the observed low-angle of the cross-bedding, only the lower portion of the dune was likely preserved (e.g. Kocurek, 1986; Mountney, 2012). The bounding surfaces between sets were likely caused by truncation from scouring along interdune corridors preceding dunes (Kocurek, 1996; Mountney and Howell, 2000). The planar appearance could indicate the dunes are two-dimensional (IE. straight-crested) or migrated at a constant rate of migration (e.g. Mesquita et al., 2021; Mountney and Howell, 2000; Rubin, 1987) or as a result of dune scour eroding down to the water table (Kocurek and Day, 2018; Kocurek and Havholm, 1993; Veiga et al., 2002). An absence of interdune deposits between cross-sets, however, would suggest the dunes were deposited in a dry, ground-water controlled aeolian system (e.g. Kocurek, 1996; Mountney, 2012).

In the light-toned sets, the uniformity, lateral persistence, and planar nature of the laminae closely resemble laminae formed by the migration of wind ripples on Earth (e.g. Ferronato et al., 2019; Kocurek and Nielson, 1986) and elsewhere on Mars (Banham et al., 2018; Grotzinger et al., 2005; Roberts et al., 2025; Yingst et al., 2023). Thin, resistant layers between the laminae create a pinstripe pattern, a characteristic feature emerging from differing permeability-related properties in inversely-graded wind ripples (Fryberger and Schenk, 1988). In wind ripples, fine silts from migrating troughs retain moisture more effectively than the coarser grains in the ripple cores, which have larger pore spaces (Fryberger and Schenk, 1988; Mountney, 2006a). This moisture retention leads to preferential cementation and the formation of thin, pin-like lines (Fryberger and Schenk, 1988; Mountney, 2006a). These pinstripe laminae have previously been identified and associated with sulfate-bearing deposits on Mars (e.g. Grotzinger et al., 2005; Roberts et al., 2025). The cm-scale, low-angle, cross-laminated sets are likely a result of reactivation of wind-ripple deposition after slight changes in wind dynamics (Kocurek, 1996; Mountney, 2006b). Due to the lateral association with the cross-bedding found in Orinoco, these light-toned sets likely originated from migration and climbing of wind ripples over slip faces of dunes (e.g. Hunter, 1977; Mountney, 2006a). The predominance of wind-ripple laminae also supports that only the apron of the dune is preserved (e.g. Kocurek, 1986).

The induration of the dark-toned laminae—that are often found truncating sets of wind-ripple strata—may have occurred after lithification through preferential diagenesis or cementation. As fines preserved in pinstripe laminae that taper concordantly across set bounding surfaces are more tightly packed, more water is retained often leading to preferential cementation and diagenesis (e.g. Carr-Crabaugh and Dunn, 1996; Fischer et al., 2012; Heidsiek et al., 2020). I interpret that this process likely formed the resistant horizons in unit A and could explain how similar the dark-toned, resistant appearance of the pinstripes in unit A are to the slightly thicker, indurated laminae that occur at the bounding surfaces

(Figure 3.3d; Figure 3.4a and b).

An irregular surface was identified to underly tabular cross-bedding with a similar appearance to unit A prior to entering Marker Band valley (Fig. 7), possibly representing an unconformity. The rock underlying this unconformity is comprised of trough cross-stratification with a similar appearance to the trough cross-stratification which dominates the clay-sulfate transition sedimentary succession (Figure 7; e.g. Edgar et al., 2024; Gupta et al., 2022; Rapin et al., 2021). Given the irregularity of the potential unconformity, the lateral extent across hundreds of metres, and how this surface marks a distinct change in cross-stratification style from tabular cross-stratification to trough cross-stratification, the base of unit A likely represents deposition following a regionally significant aeolian supersurface (e.g. Havholm and Kocurek, 1994; Kocurek, 1996) that marks the base of the layered sulfate succession.

3.4.2 Catrimani member

3.4.2.1 Unit B

Architectural elements

Unit B outcrops as a scarp-forming, ~5-m-thick, planar stratified, tabular bed, marked by a prominent sharp, planar, and laterally extensive contact with the underlying unit A. In vertical exposures, this unit maintains a consistent thickness for nearly 350 m along the western extent of Marker Band valley in Orinoco butte (Figure 3.6a), ~120 m along the northern face of Bolivar (Figure 3.6c and d), and elsewhere in vertical exposures in Marker Band valley (Supplementary figures).

Along the rover's traverse, unfavourable exposure, weathering and overprinting, and float blocks obscured fine-scale primary structures. Overprinting was most evident in the form of pervasive, rounded nodules ~3–4 mm in diameter that disrupted the visibility of finer sedimentary textures (Figure 3.7e). However, evidence of planar stratification was still preserved (Figure 3.7e). The contact between unit A and B was inferred where cross-bedding and the resistant bands of unit A were no longer visible.

Displaced blocks interpreted as unit B exhibiting a lower degree of diagenetic alteration were also encountered along the rover traverse (Figure 3.7a, Figure 3.7c). These blocks were characterised by homogenous, light-toned planar bedding composed of distinctly uniform, parallel, laterally extensive laminae with apparent thicknesses of $\sim 1 \pm 0.3$ mm (3–4 pixels thick, or approximately 0.9–1.2 mm; Figure 3.7b and B2d). Notably, these laminae did not include the fin-forming, dark-toned, resistant laminae or horizons characteristic of unit A (Figure 3.4a and b). The laminae in unit B were relatively soft, as confirmed by DRT scratches, and exhibited a bright white tone when driven over by the rover. Rounded nodular textures, ~3–4 mm in diameter, were also present but less pervasive than those observed in, in situ exposures (Figure 3.7b). Based on the light-toned appearance, planar stratigraphy, absence of resistant, dark-toned pinstripes observed in unit A, and presence of nodular textures similar to those seen in-situ within unit B, I infer these blocks as fragments from unit B.

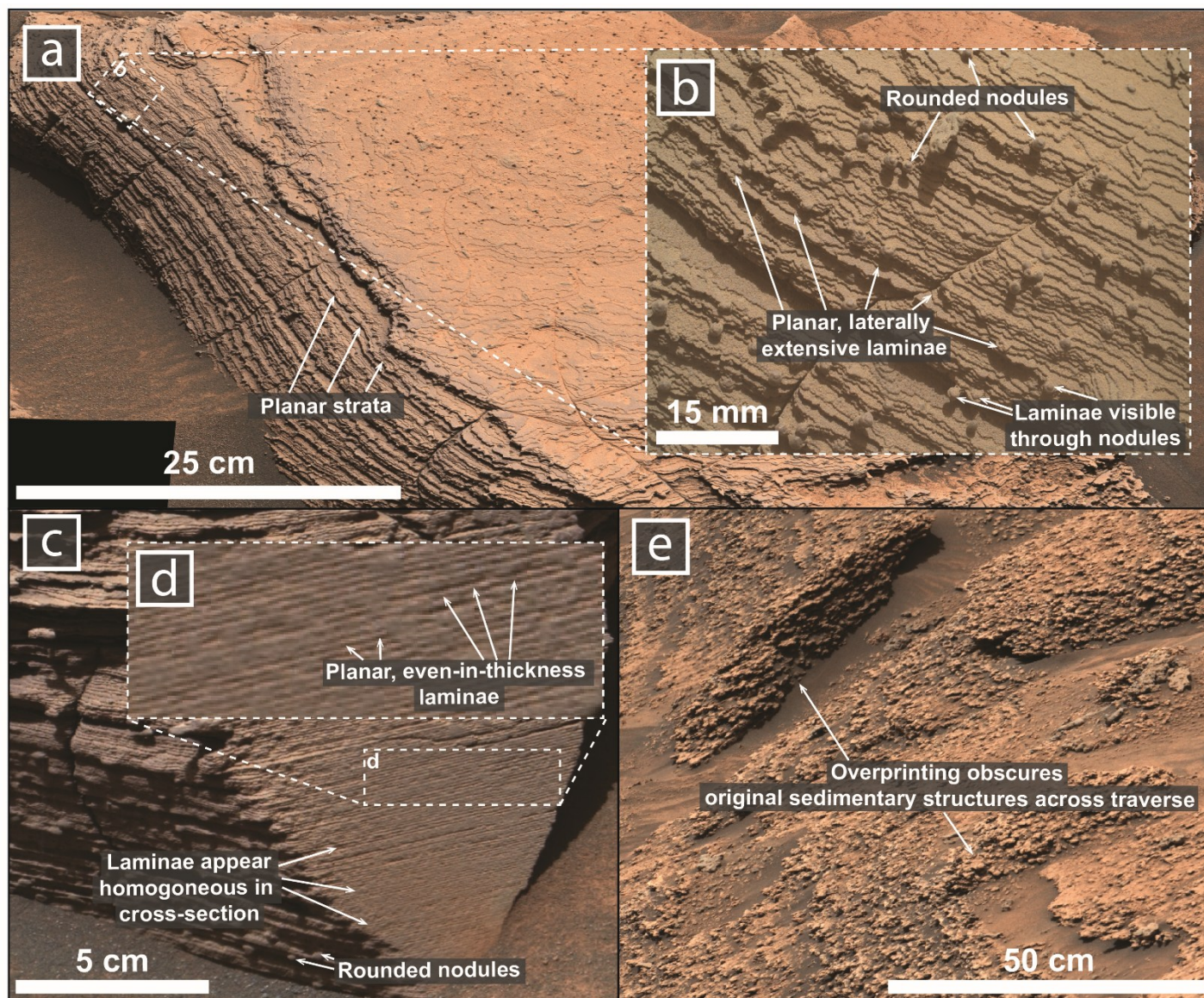


Figure 3.7: Examples of select facies characteristic of unit A as encountered across the traverse. (a) Mastcam mosaic of the light-toned, planar, laterally extensive laminae at the Canaima drill location on an out-of-situ block. (Mastcam mosaic MR_mcam103352 acquired on Sol 3705, at Sol 3703 end-of-drive location.) (b) MAHLI detail of (a). Note how planar the laminae appear in planform. (c) Mastcam mosaic showing cross-section view of the laminae on an out-of-situ block. (Mastcam mosaic MR_mcam102798 acquired on Sol 3602, at Sol 3601 end-of-drive location.) (d) Detail of (c). Note how homogeneous laminae appear in cross-section. (e) Mastcam mosaic of in-situ unit B as encountered across the rover's traverse. Extensive overprinting obscures original sedimentary structures (Mastcam mosaic MR_mcam102841 acquired on Sol 3610, at Sol 3608 end-of-drive location.) Image credits: NASA/JPL-Caltech/MSSS.

Interpretation

The planar bedding and tabular geometry of unit B, along with its direct association with underlying dune cross-bedding, would indicate either a flat-lying interdune or sandsheet depositional environment (Kocurek and Nielson, 1986; Langford and Chan, 1989; Mountney, 2006b; Mountney and Jagger, 2004; Mountney and Russell, 2009). The lateral continuity of unit B, extending for hundreds of metres both parallel and perpendicular to the wind transport direction, and the absence of any lateral transitions or contacts with dune cross-beds, favours the laterally extensive aeolian sandsheet interpretation over the interdune environment bounded by dunes interpretation (e.g. Kifumbi et al., 2022; Kocurek and Nielson, 1986; Mountney, 2006b; Rodríguez-López et al., 2010; Veiga and Spalletti, 2007). A sandsheet depositional environment would suggest limited sand supply leading to wind-driven sand accumulation and deposition through the migration of wind-ripples instead of larger dunes (Kocurek and Nielson, 1986; Loope and Simpson, 1992; Mountney, 2006a). Sand availability may have been restricted by the presence of near-surface water (e.g. Kocurek and Nielson, 1986), or slight cementation by salt (Nickling, 1984).

Nearfield observations also support sediment accumulation under sufficiently dry conditions. Although diagenesis and weathering have obscured finer-scale primary structures within in-situ outcrops, the absence of features commonly linked to wetter interdune conditions - such as symmetrical ripples and adhesion laminae (e.g. Kocurek and Nielson, 1986) - are absent. Instead, observations on the more pristine-appearing, displaced blocks attributed to unit B reveal planar, uniform in thickness, laterally extensive, ~1 mm thick laminae organised into cm-scale sets. The homogeneity and lateral continuity of the laminae are diagnostic of subcritically climbing translational strata, formed by the migration of wind-ripples (Chakraborty and Chakraborty, 2001; Fryberger et al., 1979; Fryberger and Schenk, 1988; Hunter, 1977; Kocurek and Nielson, 1986). The uniformity of laminae thickness may reflect steady winds and near-constant deposition rates (Hunter, 1977; Trewin, 1993). Alternatively, the laminae could reflect planebed lamination, formed under high wind velocities on flat, smooth surfaces (Clemmensen and Abrahamsen, 1983; Hunter, 1977; Kocurek and Dott, 1981; Ventra et al., 2017). A third possibility is that the laminae were deposited as aeolian loess; however, this interpretation is less likely due to the absence of massive bedding which is typically observed in such deposits (e.g. Pécsi, 1990).

The basal contact of unit B is interpreted as a major bounding surface. As a bounding surface is consistently identified below the Marker Band marking a ~5 m thick succession of planar strata of unit B hundreds of metres away from Marker Band valley, this major bounding surface could represent more regional scale erg deflation or contraction, and - tentatively - a supersurface, due to its lateral extent (Kocurek, 1988; Loope and Simpson, 1992; Mountney, 2006b). A period of widespread aeolian deflation could be explained by some mechanism limiting regional sand availability, which in turn promoted winds unsaturated in sediment and enhanced erosion (Kocurek, 1988; Kocurek and Day, 2018; Loope, 1985). Sandsheets are not uncommon after such supersurfaces (Kifumbi et al., 2022;

Kocurek and Nielson, 1986; Mountney, 2006b), as rising groundwater level accompanied by sediment availability at an equivalent rate can lead to promotion of aeolian sand accumulation without dune development. The flat and tabular geometry of the stratification is characteristic of sand-sheet accumulation controlled by a high-water table (e.g., Kocurek and Nielson, 1986). Alternatively, a transition into a sandsheet dominated erg marginal environment (e.g. Kocurek and Nielson, 1986) is also equally as likely.

3.4.2.2 Unit C

Architectural elements

Unit C, approximately ~0.3 m thick, outcrops as a light-toned, structureless unit with an irregular lower contact that forms a pronounced, cycloidal-like pattern against the underlying strata (Figure 3.8). Laterally, the unit appears to extend up to ~350 m across Orinoco's eastern flank (Figure 3.8) and ~120 m across the northern and southern faces of Bolivar. HiRISE images suggest this unit continues beyond Marker Band valley, with possible exposures over ~2 km away (Farrand et al., 2025). In cross-section, unit C is characterised by a mottled, light-toned, structureless material that overlays and infills frequent V-shaped wedges (~0.03 m thick, ~0.02 m width; Figures 2.10b and 2.10e). Rounded, concave-down sections of undisturbed planar stratification from unit B separate each wedge (Figure 3.8b and 10e). Although wedge spacing is not clearly uniform, these wedges recur consistently across the total extent of Orinoco. Pronounced south-dipping fractures delineate the south side of each tapered wedge, rarely exceeding a vertical extent of 0.3 m - matching the total unit thickness - and appear to cross-cut adjacent wedges (Figure 3.8c). The uppermost ~0.05 m of unit C lacks visible fractures, protrudes slightly, and shows a nodular, mottled texture (Figure 3.8b).

Across the rover's traverse, the sloped topography causes sand and detritus to mantle the overlying, indurated Marker Band (unit D), effectively concealing the ~30 cm of stratigraphy where unit C might outcrop. However, in one location, the rover encountered bedrock with a purple and light-toned mottling directly below the Amapari member (Figure Cd and Ce). MAHLI images of DRT target Mamupi (Figure 3.8e) nearby also revealed a soft, structureless, and mottled texture. Localised mottling was observed in nearfield images rover as ~5 mm diameter patches of light-toned material and ~1 mm grey-green semi-circular spots against a background that was predominantly reddish in colour (Figure 3.8e). The mottled, structureless appearance and stratigraphic context leads me to infer that this target was an exposure of unit C.

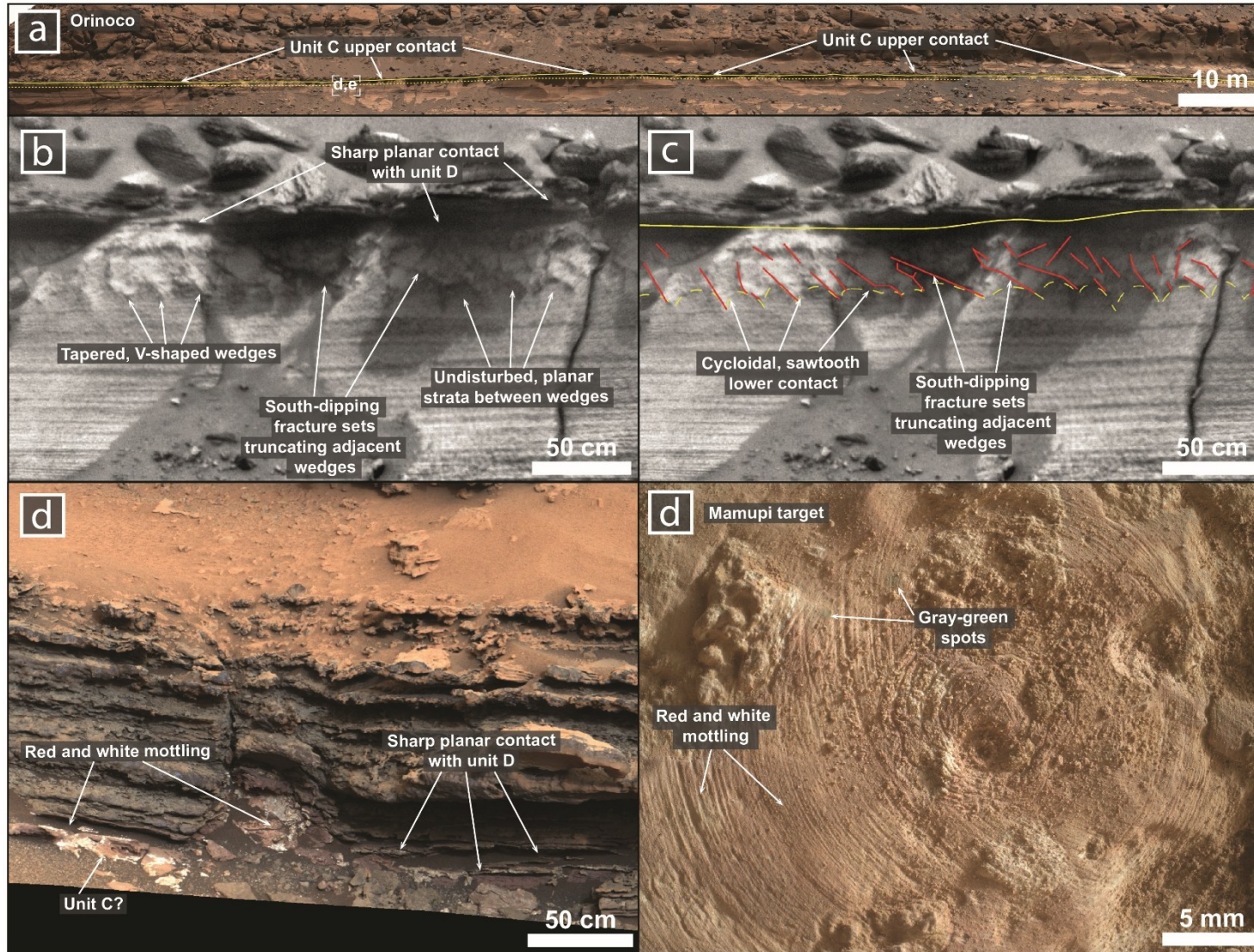


Figure 3.8: Outcrop expression and lithofacies of unit C. (a) Mastcam mosaic of the western face of Orinoco butte showing unit C, which is laterally extensive (~350 m) and has a thickness around ~0.3 m (Mastcam mosaic MR_mc102754 acquired on Sol 3596, at Sol 3594 end-of-drive location). (b) Long Distance Remote Micro-Imager (LD RMI) mosaic detail of unit C from (a), highlighting a mottled, light-toned unit underlying the indurated Amapari member. (Ccam_03625, acquired on Sol 3623, at Sol 3608 end-of-drive location). (c) Interpretation of (b) showing the unit boundaries (yellow dashed lines) and inclined fracture sets (red). (d) Mastcam mosaic of bedrock with a mottled red and white texture identified under the Amapari member encountered across the traverse (Mastcam mosaic MR_mc103019 acquired on Sol 3642, at Sol 3640 end-of-drive location). (e) A MAHLI mosaic of DRT target Mamupi acquired at same location as (d) showing a red and white mottled texture. Image credits: (a, d-e) NASA/JPL-Caltech/MSSS; (b-c) NASA/JPL/LANL/CNES/IRAP.

Interpretation

Unit C likely represents an alteration horizon. Its lateral extent - spanning hundreds of metres across multiple buttes and visible in HiRISE images over 2 km from Marker Band valley (Farrand et al., 2025) - suggests this horizon is at least locally significant. As the fracture sets are predominantly confined to unit C, penetrating only minimally into the top of unit B, and the overlying bedding in unit D is undisturbed, fracturing would likely have been temporally localised between the deposition of the sandsheets in unit B (Section 4.2.1) and the subsequent deposition of unit D. Unit B must have also been sufficiently coherent and indurated to fracture in a brittle manner without significant deformation of the original sedimentary structures (e.g. Banham et al., 2024b; Kocurek and Hunter, 1986; McMahon et al., 2017). Cross-cutting relationships, where south-dipping fractures intersect adjacent V-shaped wedges, may suggest multiple fracturing phases. Reactivation and propagation of original fracture sets could have also been caused by subsequent loading stresses (e.g. McMahon et al., 2017). The geometry of the lower contact of unit C, characterised by a cycloidal-like pattern and downward-tapering V-shaped wedges (Figure 3.8c), along with the fracture sets, is consistent with contraction or contraction-expansion processes (e.g. Kocurek and Hunter, 1986). In aeolian settings, such processes can result from wetting-drying cycles, thermal contraction, or freeze-thawing (Banham et al., 2024b; Dal Bó et al., 2010; Kocurek and Hunter, 1986; Mescolotti et al., 2019; Rodríguez-López et al., 2022).

The mottled texture, softness, and colouration identified in unit C, coupled with heterogeneous chemistry identified in the ~0.4 m of elevation below unit D (Thompson et al., 2024), likely reflects surface alteration at the time of fracture formation. This is supported by elevated Cl, the presence of hematite, and distinct multispectral variations compared to the surrounding strata (Farrand et al., 2025; Thompson et al., 2024). Farrand et al. (2025) proposed that this alteration may be attributed to wetting-drying cycles due to the association with the overlying unit D that has been interpreted as a lacustrine environment (Section 4.3.1; Gupta et al., 2023b; Mondro et al., 2025b; Weitz et al., 2023). Short-lived, wetting-drying cycles can produce wedge-shaped desiccation cracks reaching depths of 0.3 m that can be infilled with contrasting facies (e.g. Mescolotti et al., 2019). More prolonged wetting-drying cycles may lead to in-situ alteration, characterised by wedge-shaped peds and paleosol development (Ahmad, 1996; Dal Bó et al., 2010). However, formation of V-shaped wedges in both scenarios requires significant clay content (Basilici and Dal'Bó, 2010; Basma et al., 1996; Harianto et al., 2008), which is absent in unit C (e.g. Farrand et al., 2025; Thompson et al., 2024), making wetting-drying cycles an unlikely source of the fracturing.

Other contraction-expansion mechanisms that do not require clay, such as thermal contraction (Kocurek and Hunter, 1986) or even freeze-thaw cycles (Banham et al., 2024b; Benison, 2019), also align well with the cycloidal geometry of the contact and the downward-tapering wedges and their regional extent. In both cases, sediment cemented by sulfates from evaporated groundwater was proposed as the mechanism for early cohesion (Benison, 2019; Kocurek and Hunter, 1986). Tapered fractures have also been found in modern day salar flats which experience both wetting-drying cycles

and extreme temperature variations (Benison, 2019). Mg sulfate has been identified in the Catrimani member (Chipera et al., 2023; Farrand et al., 2025; Thompson et al., 2024) and early cementation of this salt may have provided the necessary cohesion for fracturing and V-shaped wedge formation in unit C. Without grain size information on these wedges, this is the preferred interpretation. Subsequent fine sediment may have been windblown or waterlain to infill and overlay these cracks and then recrystallized or diagenetically altered through wetting-drying cycles and/or temperature extremes.

3.4.3 Amapari Member

3.4.3.1 Unit D

Observations

Unit D, otherwise known as the Amapari Marker Band (extensively covered in: Gupta et al., 2023b; Mondro et al., 2025a; Mondro et al., 2025b; Thompson et al., 2024), outcrops as a tabular bed (~0.5-1 m) with a lower, dark-toned, rippled, resistant component and a recessive, light-toned, planar-laminated upper component. The contact between the lower and upper components is gradational. In Marker Band valley, the lower contact of unit D is predominantly planar, parallel to underlying bounding surfaces. Outside the study area, the lower contact appears more irregular (see Mondro et al., 2025b; Rapin et al., 2021). This unit is preserved as a protruding shelf in the buttes around Marker Band valley where it can be traced laterally for up to ~250 metres in Mastcam images of Orinoco's eastern flank. Elsewhere, this unit acts as a cap, forming a distinctive platform over buttes and valleys, including the valley west of Chenapau where the rover traversed. Orbital mapping attempts have shown that this unit can be traced for upward of 1.5 km laterally (Meyer et al., 2025; Weitz et al., 2022).

Nearfield observations of unit D show a lower facies comprised of four resistant bedsets of symmetric ripple forms (with a ~4.5 cm wavelength; Gupta et al., 2023b; Mondro et al., 2025b), that are organised as stacked or upbuilding chevrons (Gupta et al., 2023b; Mondro et al., 2025b). In plan-view, ripple crestlines are 2D, generally straight, with strikes orientated at ~120°/300° that occasionally bifurcate (Gupta et al., 2023b; Mondro et al., 2025b). The upper component of unit D is comprised of periodic, ~1 cm thick bundles of laminae (Lewis et al., 2023; Mondro et al., 2025b). These sets are planar, homogenous, laterally continuous over metres (Figure 3.9.1d and e), and overlie underlying laminae sets concordantly with no visible truncations. Laminae are often indeterminate. However, when visible, a single lamina often preserves a thickness that is smaller than 0.5 mm, ranging between 0.2 to 0.4 mm thicknesses. These laminae are dark gray in colour and have a slight lustre. Laminae sets thin upwards to 0.4 cm thick laminae at the contact with the overlying unit (Figure 3.9; (Lewis et al., 2023; Mondro et al., 2025b).

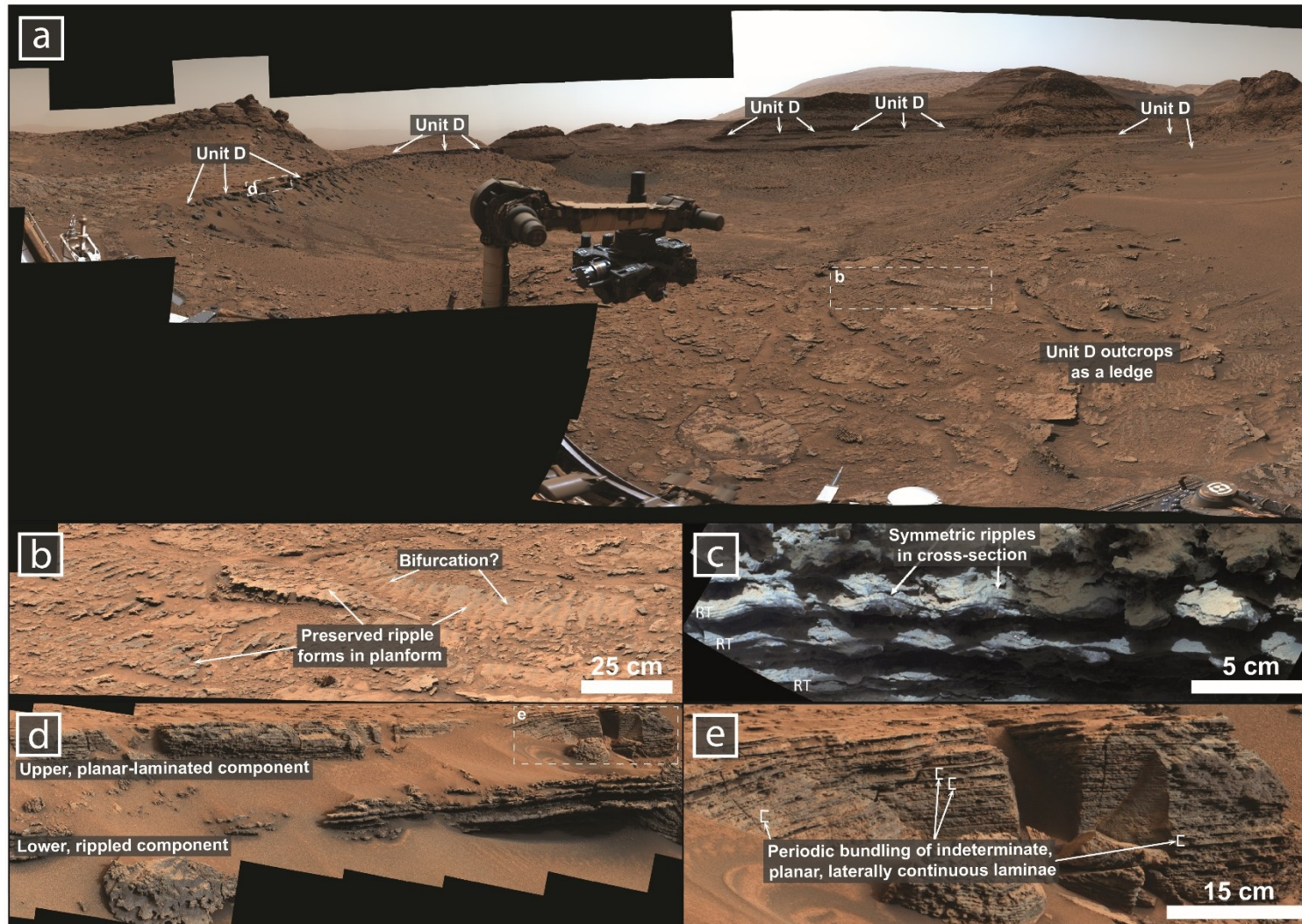


Figure 3.9: Outcrop expression and lithofacies of unit D (the Amapari member). (a) Mastcam panorama mosaic of unit D showing the lateral extent and outcrop expression as either a thin indurated ledge or bench (Mastcam mosaic MR_mcam103228 acquired on Sol 3642, at Sol 3640 end-of-drive location). (b) Detail of unit D from (a), highlighting 2D preserved ripple forms in planform view. (Mastcam mosaic MR_mcam103274 acquired on Sol 3684, at Sol 3672 end-of-drive location). (c) A MAHLI dogseye Mosaic of the symmetrical ripples that comprise the lower-component of the Amapari member. (d) Mastcam mosaic of the lower, rippled component and the upper, planar-laminated component that comprise unit D. Note the more recessed appearance of the upper component. (Mastcam mosaic MR_mcam103019 acquired on Sol 3642, at Sol 3640 end-of-drive location). (e) Detail of (d) showing the periodic bundling of the planar, laterally continuous laminae of the upper component of unit D. Image credits: NASA/JPL-Caltech/MSSS.

Interpretation

Unit D, is interpreted as a shallow lake deposit dominated by symmetrical wave ripples formed under oscillatory wave action (Mondro et al., 2025b). The presence of symmetrical ripple forms, with rounded crests and stacked chevrons, indicates sediment accretion through vertical upbuilding in response to wave-driven oscillatory flow (Gupta et al., 2023b). The short crest-to-crest wavelengths (~4.5–5 cm) suggest formation in shallow water depths, where the lake surface was open to the atmosphere and being acted upon by wind (Gupta et al., 2023b; Mondro et al., 2025b). To form wave-generated ripples, a water depth of less than ~2 m would have been required (Mondro et al., 2025a).

While the planar and laterally extensive nature of the laminae in the upper component of unit D is superficially similar to the laminae identified in units B and A, the colour, thicknesses, absence of any preserved truncation, and distinct bundling into cm-scale sets are distinct to unit D alone. These characteristics and the gradational contact with the underlying, rippled, lower component are consistent with the subaqueous settling of fines in a deepening lake (Gupta et al., 2023b). The 0.2–0.4 mm thicknesses of the laminae are also consistent with previously identified lacustrine laminae on Mars (Edgar et al., 2020). Depositional cyclicity is also inferred from the presence of the repeated bundles of laminae (Lewis et al., 2023; Rapin et al., 2025). The lateral continuity of the ripples in this unit, traceable for hundreds of metres (Mondro et al., 2025b), and as this dark-toned band is visible for over 1.5 km in orbital data may imply that this shallow lake was large, potentially rising and expanding outside of Marker Band valley.

3.4.4 Chenapau Member

3.4.4.1 Unit E

Architectural elements

Unit E outcrops as a scarp-forming, ~3.6 m thick, tabular bed of planar stratification. At the base of unit E, the contact is marked by a resistant band (~1-2 cm thick) that maintains lateral continuity for over 100 metres without interruption (Figure 3.10a, b and c), overprinting both the contact and overlying laminae and corresponding with a subtle overhang. Immediately above the resistant band, the lower component of this unit (~20 cm thick) is comprised of undulous, parallel laminae that appear to have been upwarped, forming occasional cusps (Figure 3.10c; individual cusps are up to ~10 cm in width). Resistant and recessive bands, primarily associated with these undulous laminae, are observed locally but are not consistently across all near-field outcrops. Undulous laminae transition upwards into continuous accumulations of planar, parallel laminae (~1mm thick) that are laterally extensive on a metre to decametre scale (Figure 3.10c). Up section, unit E transitions from an overhanging escarpment to a more gently sloped and recessed profile (Figure 3.10a).

An indurated, dark-toned lens (~7 m x ~0.5 m) is embedded near the uppermost extent of unit E (Figure 3.11) along the Canta detour. The lens exhibits a sharp upper and lower contact with the

surrounding bedrock. The lens (covered extensively in: Roberts *et al.*, 2025) is u-shaped, asymmetric, and exhibits a lateral pinching to the south with a wing-shaped geometry. Internally, the lens is composed of amalgamated bedsets of centimetre-scale tabular cross-bedding dipping to the northeast (Figure 3.11b and c). This feature contrasts sharply with the surrounding unit E material in both tone and resistance.

Across the rover traverse, the contact between unit D and unit E is observed by an abrupt change from planar laminae to a massive, nodular zone (Figure 3.10d and e). The nodules commonly obscure primary sedimentary structures and, at the basal contact, are organised into irregular, branching forms, which vary in size (1 – 10 mm thick; Figure 3.10e). These nodules are roughly aligned in layers and become smaller and less abundant higher in the section, where planar laminae are more apparent (Figure 3.10d).

Interpretation

I interpret that the sharp, planar, laterally extensive basal contact and the change in lithofacies marks a transition from lacustrine conditions of unit D. The steep, overhanging escarpment at the base of unit E also has a differential erosional pattern to the underlying unit D, further supporting a change in sedimentary texture such as grain size, sorting, or cementation (e.g. Edgett and Sarkar, 2021). However, the original sedimentary structures have been overprinted by diagenesis, making its precise depositional setting uncertain. Roberts *et al.* (2025) interpreted that the dark-toned lens embedded within unit E—with the asymmetric central body, wing-shaped features, and cm-scale amalgamated bedsets of tabular cross-stratification—is consistent with fluvial erosion and infill. I subsequently interpret that fluvial scouring and infill within unit E indicates that the unit must have been deposited in a subaerial environment.

At the base of unit E, diagenetic overprinting complicates direct evidence of a typical regressive sequence expected from a transition from deep lacustrine to aeolian deposition. Parallel, undulating laminae laterally transition into a massive bed comprised of branching nodular aggregates which appear superficially similar to salt-crust draping and popcorn salt aggregates from primary evaporite deposition (Goodall *et al.*, 2000). However, the planar transitions of laminae into nodules suggests that these nodules formed primarily through overprinting by disruptive crystal growth (e.g. Ahlbrandt and Fryberger, 1981). Later diagenesis likely acted to overprint original sedimentary structures. This direct switch from deep lacustrine to aeolian, with no preserved regressive sequence, indicates that wind eroded lacustrine or sabkha deposits prior to subsequent aeolian deposition, a common type of super surface called a sand-drift surface (Clemmensen and Tirsgaard, 1990; Kocurek, 1998).

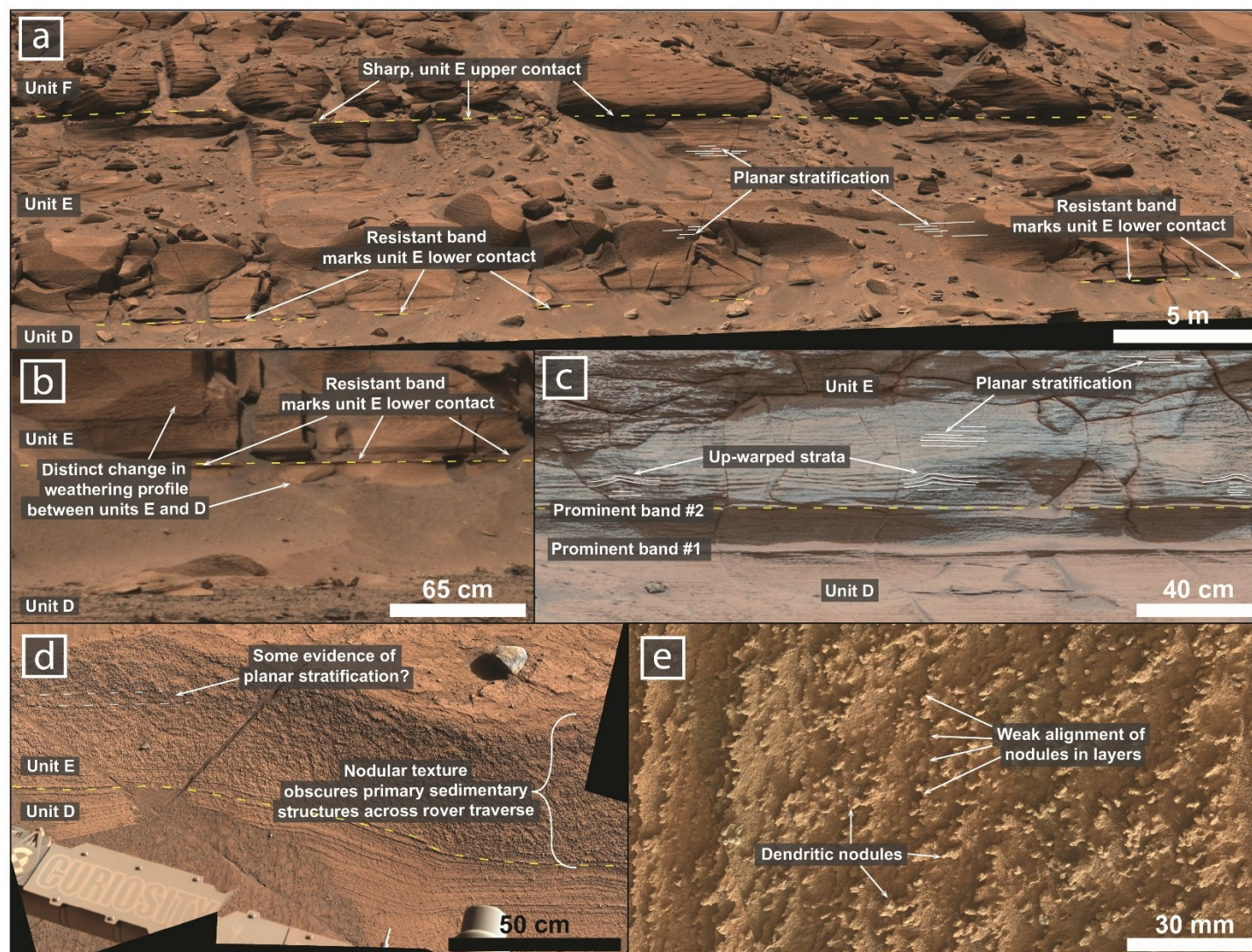


Figure 3.10: Outcrop expression and lithofacies of unit E. (a) Mastcam mosaic of the western face of Chenapau, showing the laterally extensive, sharp upper and lower contacts of unit E (mastcam mosaic MR_mcam103485 acquired on Sol 3727, at Sol 3724 end-of-drive location). (b) Detail of (a), highlighting the distinct change from a more recessed and slope forming weathering profile of the upper component of unit D to the more resistant and scarp-forming weathering profile of unit E. Note the resistant band which marks the lower contact of unit E (Mastcam mosaic MR_mcam103386 acquired on Sol 3710, at Sol 3708 end-of-drive location). (c) Mastcam mosaic of the Amapa butte showing the continuous planar stratification of unit E in cross-section. Note localised upwarping at the base and the two prominent bands which delineate the lower contact with unit D. (Mastcam mosaic MR_mcam103387 acquired on Sol3711, at Sol 3708 end-of-drive location). (d) Mastcam mosaic showing a sharp transition from unit D into a nodular texture which obscures primary sedimentary structures as encountered across the rover traverse, inferred as the contact with unit E. (Mastcam mosaic MR_mcam103736 acquired on Sol 3771, at Sol 3771 end-of-drive location). (e) MAHLI mosaic at the location of (d) showing that nodules exhibit a weak alignment into apparent layers. Image credits: NASA/JPL-Caltech/MSSS.

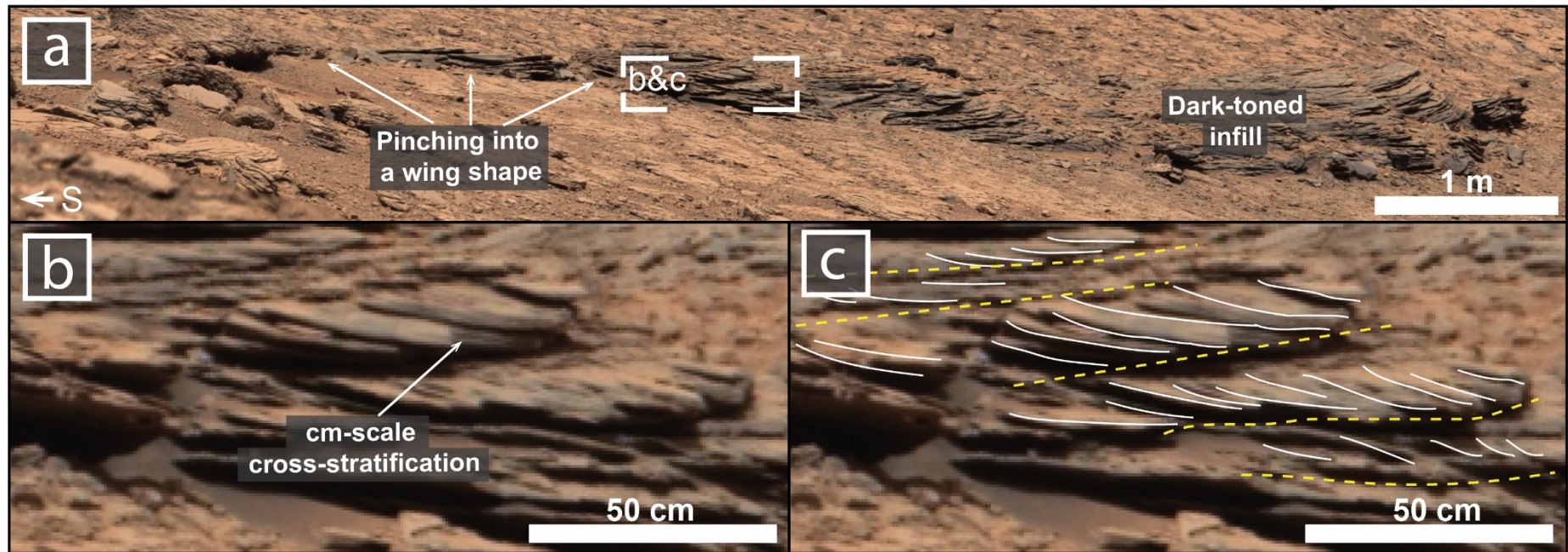


Figure 3.11: U-shaped scour embedded in unit E as encountered along the Canta detour. (a) Mastcam mosaic of the scour-and-fill structure. Note the dark-toned infill and how the scour pinches out into a wing-shape (mastcam mosaic MR_mcam103090 acquired on Sol 3653, at Sol 3651 end-of-drive location). (b) Detail showing the dark-toned scour fill. (c) Interpretation showing the cm-scale cross-bedding as described in the text. Image credits: NASA/JPL-Caltech/MSSS.

3.4.4.2 Unit F

Architectural elements

Unit F outcrops as a ~4.2 m thick, planar-stratified, scarp-forming unit with corrugated textures arranged in north-dipping ridges (Figure 3.12). The basal contact is sharp, planar, and conformable, marked by a slight overhang as unit F overlies the recessed unit E below (Figure 3.12a). Unit F is traceable for approximately ~350 metres laterally across Orinoco, ~125 metres in Chenapau, and across the valley walls (Figure 3.12). In cross-section, unit F is characterised by light- and dark-toned interbedded sets (1-4 cm thick) (Figure 3.12b) and localised corrugated textures. The corrugated textures are comprised of rounded protrusions (2-4 cm wide), relatively uniform in spacing, that are aligned along laminae.

Nearfield observations of in-situ outcrops of unit F reveal a facies composed of sets (~2–14 cm thick) of localised wavy and planar laminae (Figure 3.12c). Troughs protrude, while crests are more recessed, creating the corrugated appearance. These wavy laminae have small, mm-scale amplitudes and are irregular in both symmetry (3.58 - 2.17) and wavelength (ripple crest-to-crest spacing between ~2 to 8 cm) both laterally and vertically (Figure 3.12c). Ripple-forms appear to climb in an atypical fashion, appearing to upbuild against the stoss side and gradually increasing in wavelength up section (Figure 3.12c). Millimetre-scale nodules are interspersed throughout and often overprint original sedimentary structures.

Interpretation

The wavy laminae comprise candidate climbing ripple forms (Figure 3.12b and c). If deposited under dry aeolian conditions, the asymmetric appearance, amplitudes only a few mm thick, and wavelengths between 2 – 4 cm could be consistent with supercritically climbing aeolian wind-ripples (Clemmensen and Dam, 1993; Hunter, 1977). Thick successions of supercritically climbing wind ripples are also not unusual in terrestrial aeolian sandsheet deposits (Basilici and Dal'Bó, 2014; Simplicio and Basilici, 2015). Alternatively, a damp aeolian origin is also consistent with the observed sedimentary structures. The small-amplitude, upwind-aggradational climbing geometry, combined with irregular ripple spacing and variable symmetry index, aligns with climbing adhesion ripples, where sediment accumulates vertically on a damp ripple surface rather than migrating laterally (Clemmensen et al., 1989; Goodall et al., 2000; Jones et al., 2016; Kocurek and Fielder, 1982; Mountney and Thompson, 2002). Such adhesion deposits can form thick, laterally extensive successions in terrestrial aeolian sand sheets (Mesquita et al., 2021). The wavy laminae appear superficially similar to supercritically climbing aqueous ripples in non-channelised fluvial deposits from flooding of interdunes (Chakraborty and Chakraborty, 2001; Trewin, 1993). However, no other associated fluvial features are observed; aqueous conditions rarely persist long enough to generate more than a couple of sets of supercritically climbing ripples; the irregularity of ripple forms and spacing is inconsistent with water-lain deposition; and these wavy laminae appear distinct to aqueous ripples formed within the dark-

toned, indurated unit D, the fluvial lens in unit E, and other fluvial lenses observed in the Mirador formation (Gupta et al., 2023b; Mondro et al., 2025b). Thus, a subaqueous origin is unlikely. Consequently, I interpret unit F as a product of sustained aeolian sedimentation, where wind or adhesion processes may have controlled the accumulation of the wind ripple strata. Surficial, curvilinear ridges across the scarp faces are interpreted as erosional features formed by wind abrasion, controlled by the differential induration of laminae (Zhou et al., 2025).

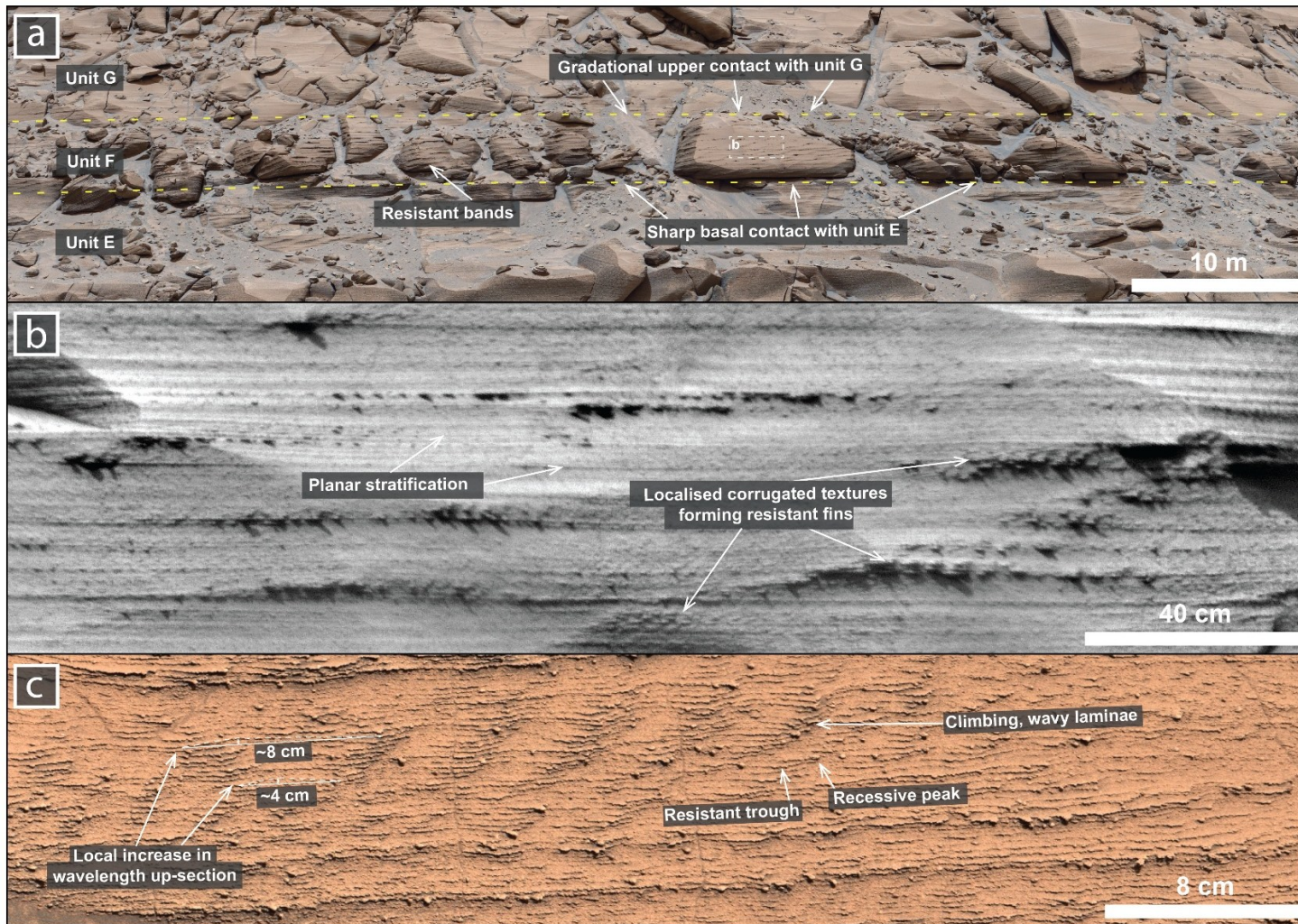


Figure 3.12: Outcrop expression and lithofacies of unit F. (a) Mastcam mosaic of Chenapau showing the laterally extensive sharp basal contact of unit F and the gradational upper contact with unit G. Note the dark-toned north-dipping bands. (Mastcam mosaic MR_mcam103485 acquired on Sol 3727, at Sol 3724 end-of-drive location). (b) LD RMI detail of (a) showing planar stratification and how the dark-toned, resistant bands are comprised of localised, resistant, corrugated textures. (ccam_02728, acquired on sol 3728, at sol 3727 end-of-drive location). (c) Localised corrugated textures as encountered across the rover's traverse. (Mastcam mosaic MR_mcam103808 acquired on Sol 3783, at Sol 3781 end-of-drive location.) Image credits: (a, c) NASA/JPL-Caltech/MSSS; (b) NASA/JPL/LANL/CNES/IRA

3.4.4.3 Unit G

Architectural elements

Unit G outcrops as a 2.7 m thick, slope-forming, light-toned tabular bed comprised of planar stratification with sporadic, discontinuous horizons of resistant fins (Figure 3.13). Unit G exhibits a gradational, basal contact from the underlying unit F marked by the gradual absence of wavy or corrugated strata.

Along the rover traverse, outcrops of unit G are distinct for their homogenous, light-toned, rounded weathering profiles (Figure 3.13). Near-field images revealed a lithofacies marked by horizontal to low-angle, planar-parallel laminae organised into sets ~40 cm in thickness but thin to ~11 cm up section. These laminae were remarkably even in thickness (~1 mm), homogenous, and laterally extensive (Figure 3.13b), similar in appearance to the planar laminae observed in unit B (Section 4.2.1). Individual laminae could be traced over several metres (Figure 3.13a). These laminae were occasionally, discordantly truncated by planar bounding surfaces at a low angle.

Discontinuous laminae of resistant “fins” or “protrusions” are sporadically distributed throughout the unit, aligned with set bounding surfaces (Figure 3.13c). These resistant laminae extend laterally over distances of 40 cm to 1 m. The fin-forming, dark-toned, indurated horizons have a slight corrugation on the mm-scale (fig 15). Equally spaced fins (~4 cm spacing) protrude for 5-10 cm from the rock and are associated with troughs in the corrugations. Rover-disturbed surfaces showed that the laminae were medium grey in colour (Figure 3.13c). Randomly distributed, millimetre-scale nodules are also common throughout this unit. Occasionally, raised “tails” emerge from protrusions, pointing upslope and to the south (e.g. Zhou et al., 2025).

Interpretation

Both the sedimentary architecture and lithofacies of unit G, with the planar stratification and the fine, even-in-thickness, parallel, laterally extensive laminae are similar in appearance to unit B (Section 4.4.2), indicating a return to deposition in an aeolian sandsheet environment. The widespread occurrence of these deposits, laterally continuous for hundreds of metres across Marker Band valley, would mean that sand supply or availability was too restricted for extensive dune construction when these deposits accumulated (Kocurek and Lancaster, 1999; Loope and Simpson, 1992; Mountney, 2006b).

I interpret that the induration and dark-colouring of the “fin-forming” resistant laminae likely occurred after lithification, through early cementation or diagenesis exploiting variations in sedimentary texture that influenced fluid flow. Their alignment along bounding surfaces implies a connection between erosive events that truncated the lamina sets and the development of these indurated horizons. The corrugated appearance of these surfaces could have resulted from wind action when the otherwise dry sand-sheet deposits were intermittently damp, where capillary action results in mm to cm-scale topography during erosion (Kocurek, 1981). Discontinuous deposition of even a small amount of fine-

clays, cements, or adherence of fine sand settling from suspension at the surface during damp or wet intervals (e.g. Chakraborty and Chakraborty, 2001; Murton et al., 2025) may have impacted the fluid retention properties, leading to preferential diagenesis or cementation at these intervals.

Tails carved into the bedrock behind the fins likely formed when these more resistant ridges shielded the adjacent laminae from upslope wind erosion, resulting in differential exhumation (e.g. Zhou et al., 2025).

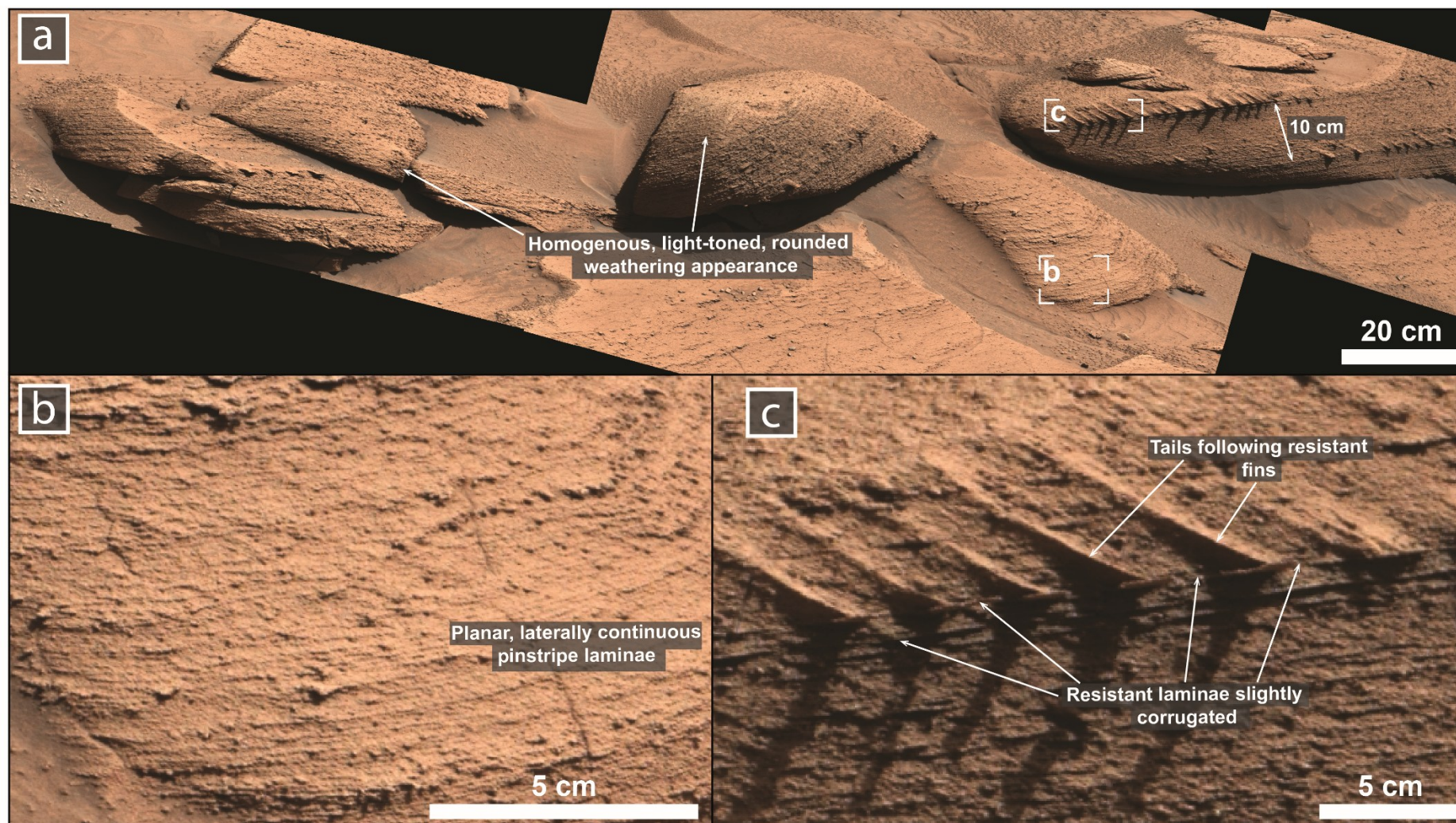


Figure 3.13: Outcrop expression and lithofacies of unit F. (a) Mastcam mosaic of unit G as encountered across the rover traverse. Note the distinctive, light-toned, rounded, homogenous appearance of the blocks. (Mastcam mosaic MR_mcaml03899 acquired on Sol 3798, at Sol 3797 end-of-drive location). (b) Detail of (a) showing the homogenous, mm-scale, planar, laterally extensive laminae which comprise unit G. (c) Detail of (a) showing how the sporadic, discontinuous, resistant, fin-forming laminae have a slight corrugation. Note the tails following the resistant fins caused by subsequent wind erosion. Image credits: NASA/JPL-Caltech/MSSS.

3.4.4.4 Unit H

Architectural elements

Unit H, ~5.3-m-thick, appears as a dark-toned, scarp-forming stratigraphic interval characterised by light-toned, recessive sets of planar laminae separated by thin indurated strata (Figure 3.14 and 17). This unit is laterally extensive, traceable for approximately 250 m in the Orinoco butte, 125 m in the Chenapau butte, and forms a cap on the first ledge in Ascent valley. The basal contact of unit H is gradational, marked by both an increase in the number of horizons of interbedded indurated laminae and their lateral extents. Truncating unit H are a set of nine distinct, broad, concave-up scour-and-fill structures (~60 to ~40 m x ~1 to 4 m) are observed. The scours are simple, symmetric, with occasional planar bases, infilled by unit I (Figure 16a; Roberts et al., 2025).

Across the traverse, unit H was easily recognised for the distinct weathering appearance of more angular blocks with interbedded recessive and thin resistant strata that give a distinctly “etched” appearance (Figure 3.14c). Nearfield images revealed that unit H is comprised of tabular sets that are between ~1 cm to 12 cm thick, consisting of planar-parallel laminae, each composed of a thicker light-toned component and a thin dark-toned component, forming a “pinstripe” pattern (Figure 3.15a). In some sections (Figure 3.15a), the laminae were heterogenous in thickness (ranging from ~1 to ~3 mm), differing from the homogenous laminae that comprise units B and G but were still laterally extensive. Thicker laminae are often associated with more etched surfaces (Figure 3.15a and b). In other sections, the laminae appeared more like the homogenous “pinstripe” laminae in units B and G with the laminae themselves exhibiting less pronounced etching (Figure 3.15b). Occasional isolated ripple structures with planar bases were also identified (Figure 3.15b and c). Overlying laminae commonly drape over the “frozen” ripple form.

Sets are often separated by indurated, dark-toned laminae. These indurated laminae appear dark-toned, flaky, and often weather to form elongated, parallel fins spaced at ~4 - 5 cm intervals. Occasionally, these appear to have a slight corrugated appearance (Figure 3.15a) and can be traced across the outcrop for up to ~3 m. In plan view, the resistant laminae appear dark-toned and comparatively smooth at a micro-metre scale. In contrast, the light-toned laminae have a rougher surface texture, composed of sub-rounded, sub-mm-scale protrusions that appear granular.

Towards the upper portion of the unit, a distinctive 1 m-thick interval contains small, disc- or “coin”-shaped fragments that spall from bedding planes (Figure 3.15d through g). This interval is predominantly composed of irregular, laterally extensive laminae (Figure 3.15d and e). Small lenses of corrugated, discontinuous laminae are also present. These corrugated laminae were more indurated and resistant and likely the source of the spalling clasts. Sets of weakly undulating and irregular cross-strata measuring 0.7–1 cm thick, dipping to the north/north-east were also identified (Figure 3.15d). Plan view observations of bedding surfaces at the clast-bearing interval show mm-scale, needle- or diamond shaped particles aligned horizontally along the bedding plane (Figure 3.15f). Displaced, rounded light-

toned blocks comprised of planar laminae with cm-scale lenses of dark-toned, planar laminae were also identified within this upper portion but their relationship with the stratigraphy is unknown.

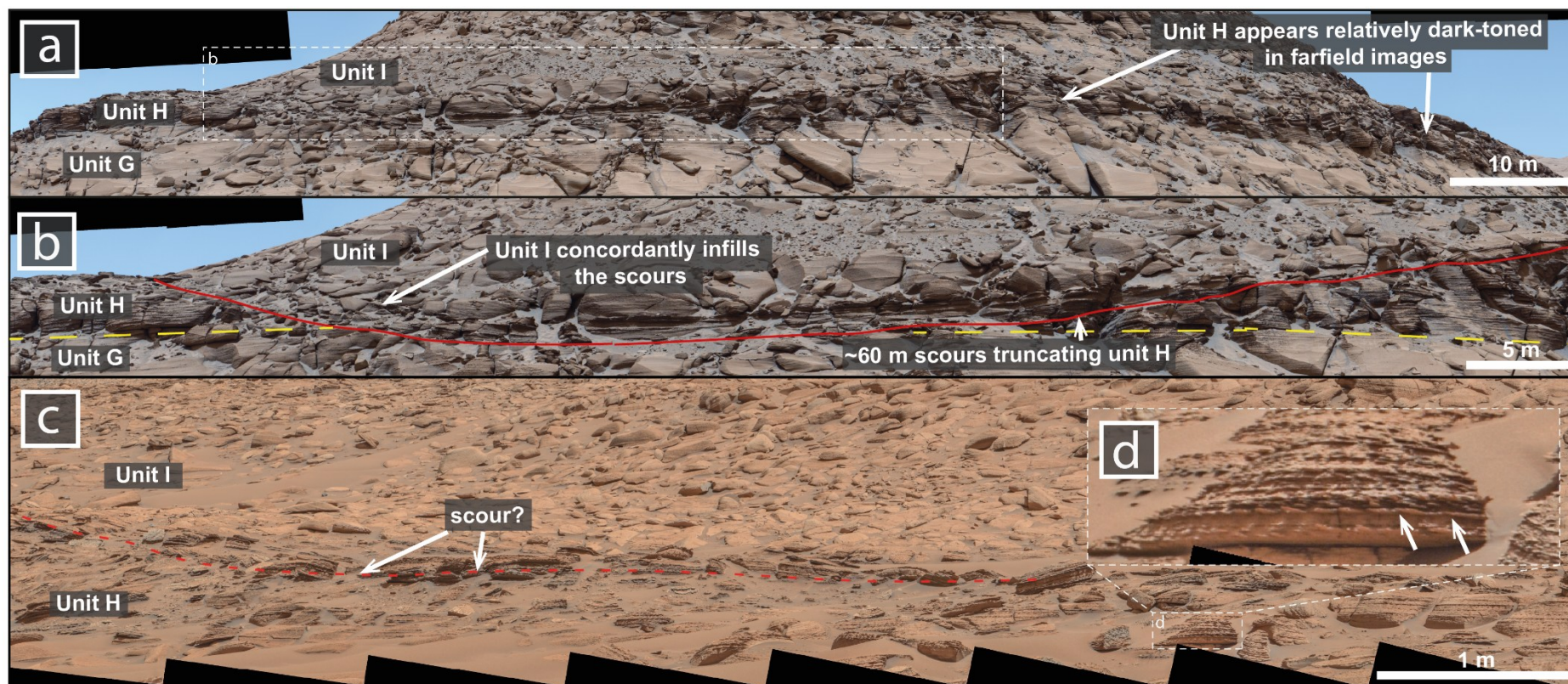


Figure 3.14: Outcrop expression and lithofacies of unit H. (a) Mastcam mosaic of the western face of Chenapau, showing the lateral extent of unit H (mastcam mosaic MR_mc103485 acquired on Sol 3727, at Sol 3724 end-of-drive location). (b) Detail of (a) showing one of the ~60 m scours truncating unit H and concordantly infilled by unit I. (c) Mastcam mosaic of unit H as encountered across the rover traverse. Note the characteristic dark-toned, interbedded resistant and recessive weathering expression and the potential scour truncating the unit H. (Mastcam mosaic MR_mc103913 acquired on Sol 3800, at Sol 3799 end-of-drive location). (d) Detail of (c) showing how resistant strata have a corrugated appearance. White arrows point out apparent micro-palaeotopography of the surfaces. Image credits: NASA/JPL-Caltech/MSSS.

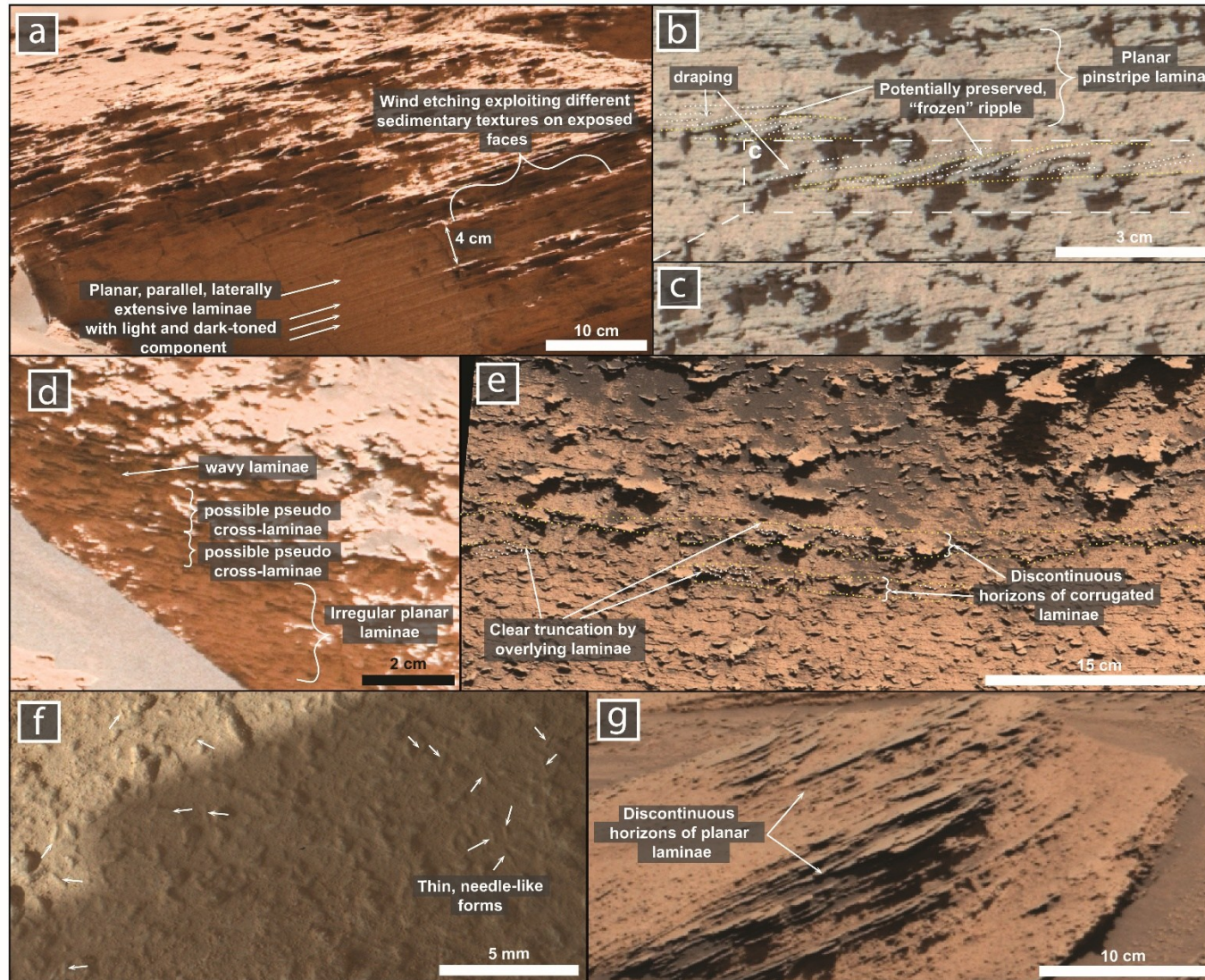


Figure 3.15: Examples of select facies characteristic of unit H as encountered across the traverse. (a) Mastcam mosaic showing resistant laminae separating cm-scale sets of recessive laminae (mastcam mosaic MR_mcam103951 acquired on Sol 3807, at Sol 3805 end-of-drive location). (b) Annotated mastcam mosaic showing sporadic “frozen” ripple forms interbedded with planar, pinstripe laminae (mastcam mosaic MR_mcam104018 acquired on Sol 3819, at Sol 3815 end-of-drive location). (c) Detail of (b) showing an unannotated frozen ripple. (d) Mastcam mosaic showing wavy and irregularly undulating laminae and candidate pseudo cross-laminae (mastcam mosaic MR_mcam103987 acquired on Sol 3812, at Sol 3810 end-of-drive location). (e) Annotated mastcam mosaic showing discontinuous lenses of corrugated laminae (mastcam mosaic MR_mcam103987 acquired on Sol 3812, at Sol 3810 end-of-drive location). (f) MAHLI mosaic showing randomly orientated thin, needle-like forms (MAHLI mosaic of target Anortosito Repartimento, taken on sol 3810.) (g) Discontinuous horizons of planar, dark-toned, resistant laminae (mastcam mosaic MR_mcam104069 acquired on Sol 3831, at Sol 3815 end-of-drive location). Image credits: NASA/JPL-Caltech/MSSS.

Interpretation

Similar to unit A (section 4.1.1), I interpret that the “pinstripe” appearance is a result of wind-ripple deposits undergoing preferential early diagenesis or cementation (Fryberger and Schenk, 1988; Mountney, 2006a). I interpret that these pinstripe laminae become distinctly “etched” on more weathered faces creating the resistant-recessive appearance, where wind exploits subtle changes in weathering resistance over long periods of time (e.g. Fryberger and Schenk, 1988). The planar-parallel appearance of the laminae and the tabular nature of the beds lead me to interpret that unit H is still deposited in a sandsheet environment.

I also interpret that there was a gradual shift in surface conditions to explain the differences with the lithofacies of unit G. The textural heterogeneity of unit H compared to unit G, however, likely represents less uniform and less sand-saturated conditions during deposition of unit H (Fryberger et al., 1979; Kocurek, 1981). Sets were also smaller marking an increase in truncation surfaces and the termination of aeolian deposition (Kocurek and Nielson, 1986) supporting less continuous wind conditions and sediment-supply limited systems (Kocurek and Lancaster, 1999; Loope and Simpson, 1992; Mountney, 2006b). Occasional isolated ripples likely represent some sort of stabilization from grain armouring, cementation, (Fryberger et al., 1979) or a rising capillary fringe “freezing” the ripple in stasis before subsequent deposition of the aeolian sand sheets (Anderson, 1990; Kocurek and Nielson, 1986).

Several processes are linked to the dark, indurated laminae. As the indurated, dark-toned, fin-forming laminae appear similar in appearance to the sporadic ones identified in unit G and retain what appears to be a slight corrugation, I also interpret that many of these bounding surfaces are linked to intermittent intervals of damp conditions and irregular deflation (Figure 3.15; Figure 3.13c; Section 4.4.3). Thick successions of interbedded damp and dry sandsheets are also common on Earth and can cause a similar “etched” resistant and non-resistant appearance (Fig. 11 in Mesquita et al., 2021; also see Chakraborty and Chakraborty, 2001; Galanin, 2021; Murton et al., 2025).

Fin-forming, resistant laminae are also associated with the thicker, heterogenous pinstripe laminae. These fin-forming laminae have a slightly different appearance for their apparent planar form and less regularly spaced fins (Figure 3.15). These pinstripe laminae in unit H also appear different to the pinstripe laminae in units B and G, leading me to interpret that a slight change in grain provenance, organisation, or packing in the wind-ripple deposits to cause more pronounced preferential cementation at these intervals (Banham et al., 2024a; Besly et al., 2018; Fryberger and Schenk, 1988).

In the uppermost interval of unit H, the irregular, wavy geometry of the pseudo-cross-laminations closely resembles climbing adhesion ripples (Goodall et al., 2000; Hunter, 1973; Kocurek and Fielder, 1982). The needle/diamond shaped features observed in the MAHLI images resemble euhedral sulfate crystals that have a platy and needle-like habit previously identified on Mars before (Kah et al., 2018; Schieber et al., 2022). While I cannot determine whether these crystals were detrital or synsedimentary interstitial crystals growing amongst pore spaces, both would indicate crystallization of sulfates likely

through evaporation at the surface or near surface when the sediment was deposited (Cohen et al., 2022; Kocurek and Nielson, 1986). Discontinuous lenses of dark-toned laminae may also support brief ponding (Chakraborty and Chakraborty, 2001). Put together, I tentatively infer that these features may represent a damp, or salt-cemented sabkha-like environment at the top of unit H.

The concave-up scour-and-fill structures that truncate unit H are interpreted as blowouts (further described in Roberts et al., 2025) formed by episodes of localised deflation by strong, sediment unsaturated winds (Hesp, 2002; Hugenholtz and Wolfe, 2006, 2009). Similar sized and shaped ‘saucer’ blowouts are often associated with sabkha environments, where surficial salt cements can act to stabilize the surface (Hesp, 2002). The planar base observed in one of these scours may indicate deflation down to the water table/salt-cemented surface (Roberts et al., 2025).

3.4.4.5 Unit I

Architectural elements

Unit I, with a thickness of ~4.9 m, is a light-toned, slope-forming interval characterised by planar stratification appearing similar to unit G (Figure 3.16). The basal contact with unit H is sharp, particularly where the unit drapes across the embedded scours (Figure 3.14a; 18a). Unit I is laterally extensive, continuous for approximately 125 m in the Chenapau butte and 250 m in the Orinoco butte and appears homogenous in cross-section.

Along the rover’s traverse, unit I is observed to be gently sloped, with weathering commonly forming rounded, smooth, light-toned blocks, similar to the weathering patterns observed in unit G and a distinct contrast to the “etched” underlying unit H (Figure 3.16a and b). However, unlike unit G, unit I lacks the discontinuous horizons of protrusions or fins, except for indurated strata in the palaeo-topographic lows at the centres of the scours. The lithofacies of unit I is characterized by planar, homogenous, and laterally extensive laminae similar to unit G (Figure 3.16c and d). Like unit G, individual laminae - traced laterally for ~3 m across outcrops - demonstrate planar geometry with no thickness variation. As a MAHLI dog’s eye was taken of this unit, detailed measurements of the laminae could be made showing a consistent laminae thickness, with an average thickness of ~0.76 mm and a standard deviation of 0.058 mm (Figure 3.16d). Like unit G, sporadic mm-scale nodules are also common in this unit.

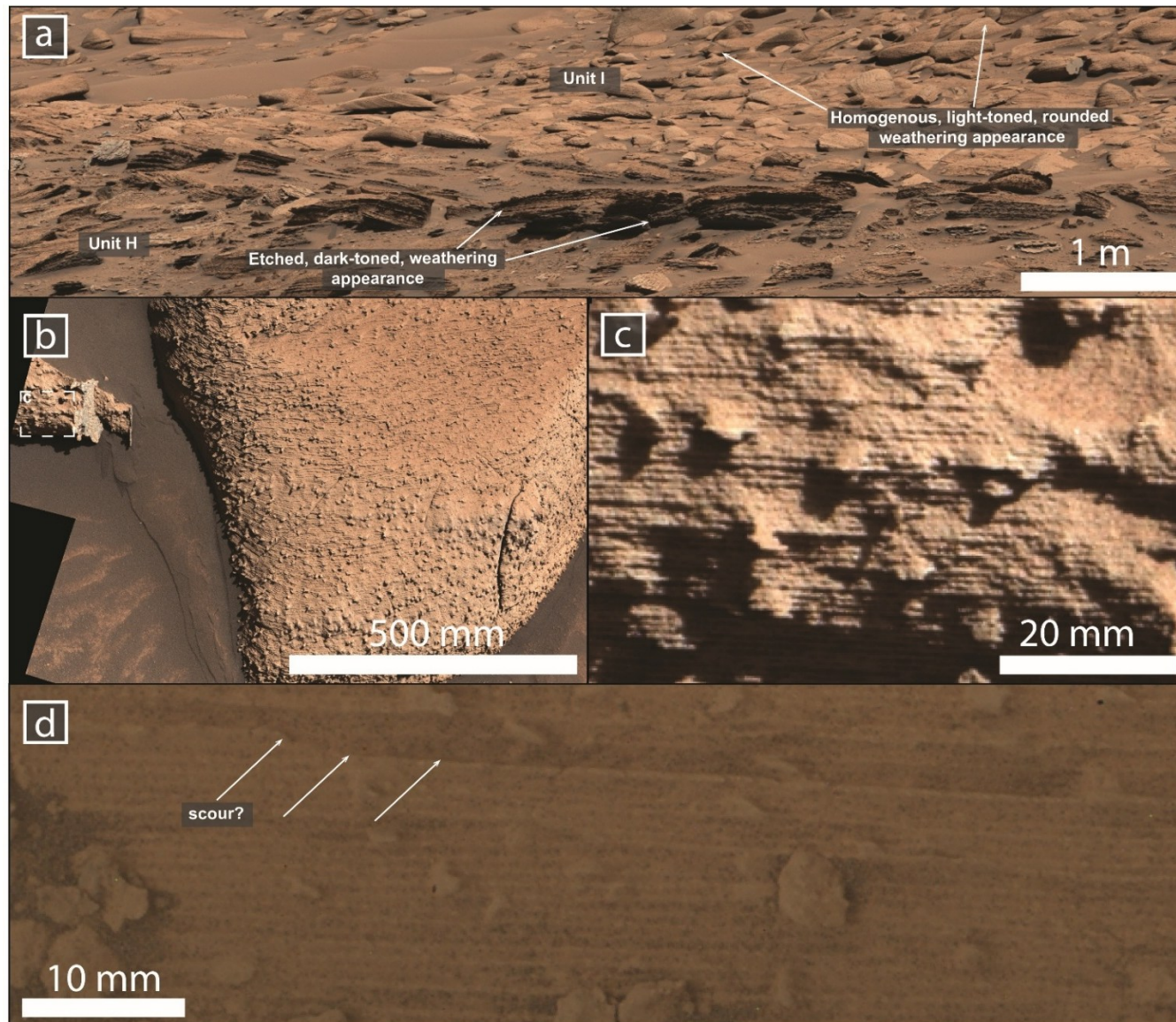


Figure 3.16: Outcrop expression and lithofacies of unit I. (a) Mastcam mosaic showing the difference in weathering appearance of unit I and unit H. Note how unit I appears homogenous, light-toned and weathers as rounded blocks compared to the dark-toned, etched weathered appearance of the underlying unit H. (Mastcam mosaic MR_mcam103913 acquired on Sol 3800, at Sol 3799 end-of-drive location.) (b) Mastcam mosaic showing the light-toned and homogenous appearance of unit H. (Mastcam mosaic MR_mcam104131 acquired on Sol 3845, at Sol 3843 end-of-drive location.) (c) Detail of (b) showing how the light-toned, homogenous appearance is caused by fine, equal-in-thickness, homogenous laminae. (d) MAHLI dogseye mosaic showing the homogenous, equal-in-thickness laminae. Note a potential ~cm scale scour that appears to truncate the laminae. (Kastria Spring target, 25 cm stand off.) Image credits: NASA/JPL-Caltech/MSSS

Interpretation

The planar-parallel, homogenous laminae lithofacies arranged into tabular, planar sets observed in unit I are interpreted to represent the same lithofacies present in units B and G. I therefore interpret that unit I also represents aeolian sandsheet deposition under continuous wind conditions and a consistent sediment supply (Section 4.4.2; 4.4.7). The observed concordance of the strata within unit I as they infill the underlying blowout scours that truncate unit H (Roberts et al., 2025), suggests that these deposits accumulated and draped across pre-existing palaeotopographic relief - an observation consistent with accumulations of ancient, terrestrial sandsheets (e.g. Langford et al., 2008). Concordant, isolated, cm-scale scours could also suggest minor erosion by either strong winds (Hesp and Hyde, 1996; Schwan, 1988) or local turbulence (Grotzinger et al., 2005) concurrent with sandsheet deposition.

3.4.4.6 Unit J*Architectural elements*

Unit J, ~7.3 m thick, is a dark-toned, scarp-forming stratigraphic interval, characterized by interbedded recessive and thin resistant strata sets similar in appearance to unit H. The basal contact of unit J is gradational with the underlying unit I and defined by an increased presence of laterally continuous resistant strata. Unit J extends across ~250 m of Orinoco and ~125 m of Chenapau. One scour (~60 m wide, ~5 m thick) is observed in Orinoco, truncating the planar stratification of unit J. This scour is also symmetric and infilled with the light-toned, planar strata characteristic of unit K (Figure 3.17a).

Like unit H, unit J was easily recognised across the traverse for the distinct weathering pattern that formed angular blocks with an etched appearance similar to unit H (Figure 3.17a, b; Figure 3.18a and b). Unit J was also comprised of tabular sets of recessive light-toned laminae separated by discontinuous resistant fin-forming laminae that occasionally appear corrugated (Figure 3.17b) and occasional “frozen ripples” (Figure 3.17c). Unlike unit H, unit J exhibits occasional scours truncating the strata (Figure 3.17b).

The light-toned, recessive sets are comprised of planar-parallel, laterally extensive laminae that have a pinstripe pattern (Figure 3.18a and b). Like unit H, these laminae are heterogeneous in thickness but are not associated with an etched appearance. The dark-toned laminae separating the sets correspond to bounding surfaces that truncate the light-toned laminae (Figure 3.20a). The dark-toned resistant laminae have a variety of textures. Some have gently irregular and undulating surfaces (Figure 3.18c), others form parallel, finger-like ridges that occasionally display a dendritic pattern, in one case associated with raised, dark-toned polygonal ridges (Figure 3.18d). Other surfaces appear to drape across U-shaped grooves in the light-toned recessive laminae (Figure 3.18E). Similar to unit H, the resistant laminae appear smoother on a micro-metre scale and lack the distinctive granular or nodular appearance.

Interpretation

The resistant-recessive, planar-parallel, heterogenous laminated sets observed in unit J resemble the lithofacies observed in unit H (Section 4.4.8). Similarly, I interpret unit J was deposited in a similar aeolian sandsheet environment as unit H characterised by non-uniform winds undersaturated in sand.

The mm-scale corrugated and irregular appearance of the dark-toned laminae in plan-view appears superficially similar to adhesion surfaces in plan-view (Basilici et al., 2021; Kocurek and Fielder, 1982). The finger-like, dendritic, dark-toned ridges and draping appear superficially similar to textures/facies that result from brief ponding during termination of aeolian deposition (e.g. Chakraborty and Chakraborty, 2001; Kocurek and Nielson, 1986). The polygonal ridges of light-toned material over the dark-toned, indurated laminae (Figure 3.18d) appear similar to other polygonal patterns identified in the clay-sulfate transition attributed to either desiccation (Rapin et al., 2023a) or diagenesis post-dating lithification (Seeger and Grotzinger, 2024).

Unit J also shows m-scale cross-bedding which may represent migration of small, isolated dunes common in some sandsheet environments (Kocurek and Nielson, 1986). The large, ~63-m-wide scour that truncates unit J resembles the scours identified truncating the strata within unit H and I similarly interpret that these represent localised deflation from palaeo-blowouts.

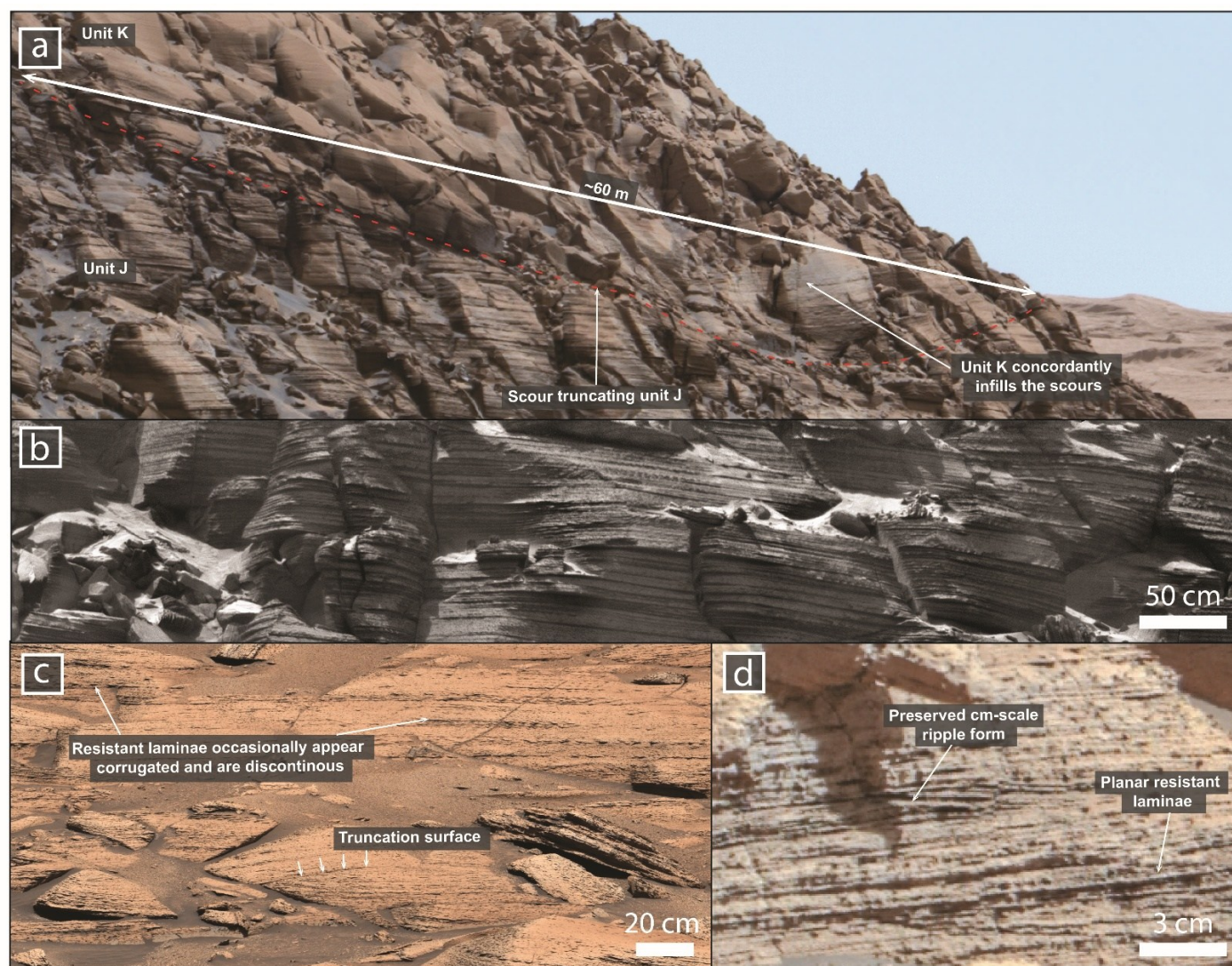


Figure 3.17: Outcrop expression and lithofacies of unit J. (a) Mastcam mosaic showing the dark-toned, planar lithofacies of unit J within the western face of Chenapau. Note the ~60 m scour. (mastcam mosaic MR_mcam103485 acquired on Sol 3727, at Sol 3724 end-of-drive location). (b) LD RMI showing how the dark-toned, planar-bedded lithofacies is comprised interbedded resistant and recessive strata (ccam 02737, acquired on sol 3737, at sol 3735 end-of-drive location). (c) Interbedded resistant and recessive packages as encountered along the rover traverse. Note how the resistant laminae occasionally appear corrugated and discontinuous and the truncation surface. (mastcam mosaic MR_mcam104297 acquired on Sol 3871, at Sol 3870 end-of-drive location). (d) Preserved cm-scale ripple form identified in one of the sets (mastcam mosaic MR_mcam104379 acquired on Sol 3885, at Sol 3880 end-of-drive location). Image credits: (a, c-d) NASA/JPL-Caltech/MSSS; (b) NASA/JPL/LANL/CNES/IRAP.

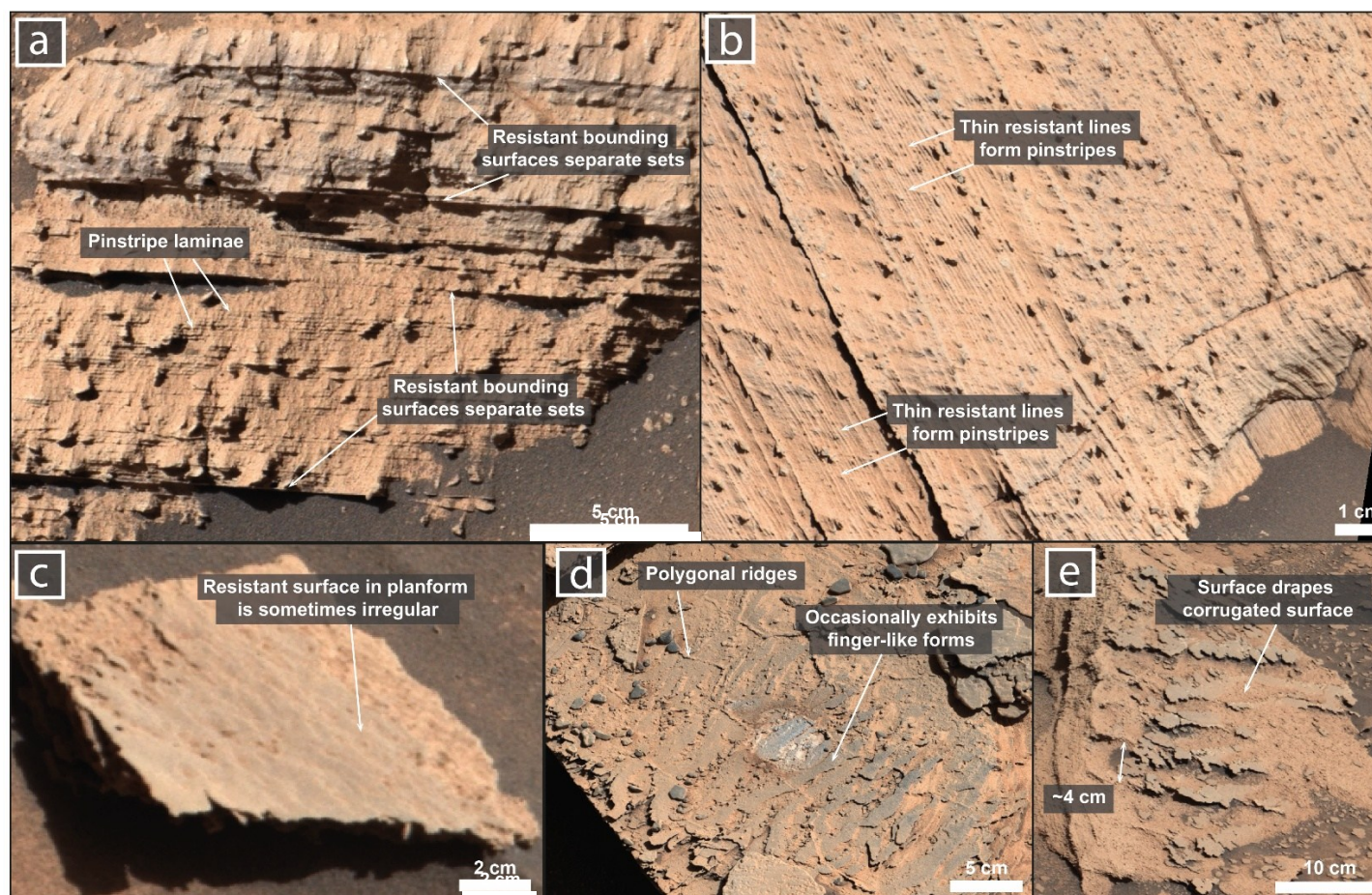


Figure 3.18: Examples of select facies characteristic of unit J as encountered across the traverse. (a) Mastcam mosaic of the light-toned, planar, laterally extensive laminae separated by resistant bounding surfaces. (Mastcam mosaic MR_mcam104278 acquired on Sol 3867, at Sol 3865 end-of-drive location.) (b) Mastcam mosaic showing the pinstripe pattern of the light-toned laminae. (Mastcam mosaic MR_mcam104420 acquired on Sol 3892.) (c-e) Mastcam mosaics showing planform views of the various textures exhibited by the resistant bounding surfaces. (c) Surface with a gently irregular and undulating topography on the mm to cm-scale (mastcam mosaic MR_mcam104328 acquired on Sol 3876, at Sol 3873 end-of-drive location.) (d) Surface comprised of parallel ridges with a dendritic pattern. Note the overlying light-toned, polygonal ridges. (Mastcam mosaic MR_mcam104461 acquired on Sol 3898, at Sol 3897 end-of-drive location.) (e) Surface comprised of u-shaped grooves truncating light-toned laminae and draped by the dark-toned, resistant surface. (Mastcam mosaic MR_mcam104403 acquired on Sol 3889, at Sol 3887 end-of-drive location.) Image credits: NASA/JPL-Caltech/MSSS.

3.4.4.7 Unit K

Architectural elements

Unit K, ~12.9 m thick, outcrops as a light-toned, slope-forming stratigraphic interval with uniform, planar stratification (Figure 3.21). Unit K marks the top of the stratigraphic succession observed in Orinoco and Chenapau, draping across a sharp, planar contact with the underlying unit J and associated scour (Figure 3.18a).

Across the traverse, unit K outcrops as light-toned, rounded, smooth blocks similar to units G and I. The lithofacies is also similar, characterised by light-toned, laterally extensive, planar, even-in-thickness laminae. While poor outcrop exposure prevented more detailed measurements of these laminae, they were consistently <1 mm where measured (Figure I.4.7) and appeared homogenous. Similar mm-scale nodules were also observed here.

Interpretation

The planar-parallel, homogenous laminae lithofacies observed in unit K are similar to the lithofacies observed in units B, G, and I (Section 4.4.2; 4.4.7; 4.4.9) and represent aeolian deposition in an aeolian sandsheet environment. The uniformity of the laminae likely reflects steady winds and near-constant deposition rates.



Figure 3.19: Example of unit K lithofacies as encountered across the traverse. (Mastcam mosaic MR_mcaml04581 acquired on Sol 3914, at Sol 3912 end-of-drive location.) Image credits: NASA/JPL-Caltech/MSSS.

3.5 Discussion

3.5.1 What was the paleoenvironmental setting?

Combined sedimentary facies and architectural-element analysis indicate that the layered sulfate-bearing strata are formed by depositional processes influenced by both wind and water. Here, I discuss the possible depositional processes responsible for forming the ~100-m-thick succession exposed within Marker Band valley, arguing for a predominantly aeolian depositional model influenced by near surface water.

3.5.1.1 Establishing an aeolian paleoenvironment

The tabular, metre-scale cross-bedding with asymptotic bases identified in the Contigo member (Figure 3.3) account for half of the total stratigraphic succession exposed in Marker Band valley. This lithofacies appears similar to cross-bedding formed by migrating simple dunes. Both fluvial and aeolian dunes can exhibit metre-scale, low-angle cross-bedding with asymptotic bases (Collinson and Mountney, 2019). However, several architectural features along with other contextual facies observations lead me to prefer an aeolian depositional interpretation. The broad, sweeping, asymptotic downlapping relationships of the cross-sets onto the underlying bounding surfaces (Figure 3.3; Section 4.1) are characteristic of dune apron accumulation from reversing winds, wind-flow detachment, or sand suspension over dune crests (Brookfield, 1977; Clemmensen, 1988; Kocurek et al., 2007; Kocurek and Dott, 1981; Walker and Middleton, 1977). There is also a notable absence of other fluvial architecture such as downcutting erosional surfaces expected from fluvial cross-beds or paleochannel forms.

One of the more compelling features supporting an aeolian origin for unit A is the prevalence of pinstripe laminae (Sections 4.1, 4.2.1, 4.4). These laminae are planar, laterally extensive, and exhibit thin, dark-toned (Figure 3.15a) or indurated (Figure 3.7d; Figure 3.16d) lines closely resembling a pinstripe pattern. *Pinstripe laminae* are a phenomenon directly attributed to inverse grading in wind ripples (Fryberger and Schenk, 1988; Mountney, 2006a). In wind-ripple deposition, fine particles accumulate in ripple troughs, shielded by coarser grains during ripple migration, producing distinctive thin lines that create this diagnostic pinstripe pattern preserved in the rock record (Fryberger and Schenk, 1988; Mountney, 2006a). Finer particles can retain moisture, undergoing preferential early cementation leading to the pinstripes becoming indurated (Fryberger and Schenk, 1988). While inverse grading is difficult to confirm on Mars as finer particles are smaller than image resolution limits (e.g. Banham et al., 2022; Stack et al., 2019), I can use the pinstripe pattern to infer the presence of laminations produced by aeolian wind ripples (Roberts et al., 2025), supporting an aeolian interpretation for unit A.

The planar bedded lithofacies identified in the Catrimani and Chenapau members account for 50% of the total stratigraphic succession exposed in Marker Band valley. Facies and architectural element analysis demonstrate shared lithofacies within the planar bedded units, indicative of the operation of broadly similar sedimentary processes. These lithofacies can broadly be subdivided into light-toned (Catrimani member: unit B, and the Chenapau member: units G, I, and K) and dark-toned, planar bedded lithofacies (Chenapau member: units H and J).

The light-toned, planar-bedded units contain highly uniform, planar-parallel laminae. This uniformity is often diagnostic of subcritically climbing translational strata, formed by the migration of wind-ripples (Chakraborty and Chakraborty, 2001; Fryberger et al., 1979; Fryberger and Schenk, 1988; Hunter, 1977; Kocurek and Dott, 1981; Kocurek and Nielson, 1986). Their arrangement into broadly

planar-bedded sets implies accumulation over a planar surface, consistent with an aeolian sand-sheet environment (Kocurek and Nielson, 1986). The uniform thickness of laminae may reflect both the regular spacing of wind ripples in a given area at one time (Hunter, 1977; Trewin, 1993) - as observed on Mars (Durán et al., 2014) - and steady winds with near-constant deposition rates (Hunter, 1977; Trewin, 1993). While some units exhibit clear evidence of a pinstripe pattern (Unit I; Section 4.9), pinstripe laminae are not as easily recognised due to either poor outcrop exposure, different fluid-retention properties (Fryberger and Schenk, 1988), or a uniform grain size (Collinson and Mountney, 2019). Alternatively, some laminae could reflect aeolian planebed lamination, which forms under high wind velocities on flat, smooth surfaces (Clemmensen and Abrahamsen, 1983; Hunter, 1977; Kocurek and Dott, 1981; Ventra et al., 2017). The small, concordant scours identified in unit I may support this interpretation, as similar features have been attributed to strong wind activity in aeolian plane-bed deposits (Schwan, 1988).

The dark-toned, planar-bedded lithofacies also contain pinstripe laminae (Section 4.1.1), suggesting continued aeolian sand-sheet deposition. Broad, u-shaped scours truncate the dark-toned, planar-bedded lithofacies and likely represent erosion by intense, localised wind deflation (e.g. Roberts et al., 2025). There is also evidence that this lithofacies may contain occurrences of a damp depositional environment. Unit H (Section 20) contains potential adhesion ripples – which form when sediment sticks to a slightly damp surface (Section 4.4.4; e.g. Goodall et al., 2000; Hunter, 1973; Kocurek and Fielder, 1982) – and needle/diamond shaped features resembling euhedral sulfate crystals (Mees et al., 2012) - suggesting syndepositional evaporite crystals that were either reworked or formed near the surface (Cohen et al., 2022; Kah et al., 2018; Kocurek and Nielson, 1986; Schieber et al., 2022).

3.5.1.2 Evidence for episodic surface water

Mondro et al. (2025b) showed how symmetric ripples in the Amapari member represent a shallow lake and how the transition into repeated bundles of planar laminae in the uppermost section of this member likely represents a deepening lacustrine environment. As wind deflation can erode substantial lacustrine deposits (Clemmensen and Tirsgaard, 1990; Kocurek, 1988), the abrupt transition into aeolian sand-sheets in the Chenapau member after the Amapari member is therefore inferred to have occurred after some amount of wind deflation, likely indicating that the preserved ~1 m thick unit is only a small portion of the original lacustrine deposits.

Further evidence for surface water is identified in the dark-toned lens embedded within unit E of the Chenapau member. A lithofacies comprised of cm-scale cross-bedding and a wing-shaped architectural feature are characteristic of fluvial dunes in a shallow stream (Roberts et al., 2025). This fluvial lens signifies that surface water was also periodically present during the accumulation of the aeolian sand-sheets in the Chenapau member.

3.5.1.3 What were the controls on deposition and what caused the thick succession of sandsheets?

The planar bedded lithofacies represent an aeolian sandsheet deposit approximately 60 m thick - substantially thicker than typical sandsheets on Earth which rarely exceed ~20 m (Mountney, 2006a), with the thickest known succession reaching ~50 m thick (Murton et al., 2025; Simplicio and Basilici, 2015). Accumulating and preserving such an unusually thick deposit of sandsheets in the Chenapau member would have required a sustained combination of conditions. First, net sediment input into the system must have occurred (Kocurek and Havholm, 1993). Second, some mechanism must have limited the amount of loose sediment available, suppressing the ability of dunes to form (Kocurek and Nielson, 1986; Simplicio and Basilici, 2015). Third, the deposit must have been protected from erosion and reworking at the surface to allow for continued vertical aggradation (Kocurek and Havholm, 1993). These conditions could have been satisfied by a number of factors including a rising water table, periodic flooding, surface cementation, coarse grain sizes or ice (Basilici and Dal'Bó, 2010; Bristow et al., 2010; Chakraborty and Chakraborty, 2001; Clemmensen and Dam, 1993; Lea and Waythomas, 1990; Simplicio and Basilici, 2015).

Water would likely have played a significant role in both the accumulation and preservation of the aeolian sandsheets. Evidence for periodic surface water – in the form of the Amapari member (Gupta et al., 2023b; Mondro et al., 2025b) and the ephemeral stream in unit E – suggests deposition occurred at or close to the water table (e.g. Mountney and Russell, 2009). In aeolian sandsheets, a high water table promotes a flat depositional surface ideal for the accumulation of highly planar and parallel wind-ripple laminae (Kocurek and Nielson, 1986), closely resembling those in the planar-bedded lithofacies of the Catrimani and Chenapau members. Adhesion ripples in unit H further suggest that the water table or its capillary fringe intersected the Martian surface, dampening the sediment enough for adhesion of sediments in transport (Clemmensen et al., 1989; Goodall et al., 2000; Grotzinger et al., 2005; Jones et al., 2016). Since damp sand has a much higher threshold velocity for entrainment due to capillary tension, a high or rising water table is an effective mechanism that can continuously limit the quantity of entrained sand, suppressing dune formation and promoting sand sheet accretion (Kocurek and Nielson, 1986). Lenticular crystal forms in unit H may also indicate syndepositional gypsum precipitation at or near the surface (Cohen et al., 2022; Kah et al., 2018; Kocurek and Nielson, 1986; Schieber et al., 2022), potentially acting as a cement and further increasing the threshold velocity required for sediment mobilisation.

3.5.1.4 What caused the dark- and light-toned appearing stratigraphic intervals?

The vertical variations between the dark-toned, planar bedded lithofacies and the light-toned, planar bedded lithofacies may be governed by variations in water table rise, surface cementation, wind energy, and the depositional rate. Fryberger et al. (1979) and Schwan (1988) showed that lithofacies from

aeolian sandsheets deposited under steady wind regimes and a consistent sand supply can differ from those deposited under more variable conditions (also see Kocurek, 1981). The light-toned, planar bedded lithofacies of the Catrimani and Chenapau members are texturally homogenous with the planar-parallel, uniform laminae and rare truncation, consistent with deposition under continuous wind conditions with readily available sand, comparable terrestrial accumulations have been hypothesised to reach similar vertical thicknesses to unit I from a single storm event (Fryberger et al., 1979; Lea and Waythomas, 1990; Schwan, 1988). Deposition under these conditions would have required a continuously rising water table to account for their accumulation (Basilici et al., 2021; Havholm and Kocurek, 1994; Kifumbi et al., 2022; Mountney, 2006b; Mountney and Jagger, 2004), but the rate of deposition was likely sufficiently high enough that the water table never intersected the surface to cause wet or damp surface conditions during the formation of the light-toned, planar-bedded units but not long enough to mobilise sediment into larger bedforms.

The dark-toned, planar bedded units, however, appear more texturally heterogeneous and record abundant truncation surfaces, which could perhaps be attributed to more varying wind conditions and less available sand (Fryberger et al., 1979; Kocurek, 1981; Lea and Waythomas, 1990; Mountney, 2006b; Schwan, 1988). As the dark-toned, planar bedded lithofacies contains evidence for potential adhesion ripples, potential syndepositional gypsum precipitation, and because of the association with the lacustrine Amapari member and fluvial lens, it is likely that episodic rises in surface water and surface cementation reduced the availability of loose sediment, leading to sediment bypass or even erosion producing the large number of bounding surfaces (e.g. Kocurek and Havholm, 1993). Rare preservation of dune forms—as seen in unit J—are also not uncommon in this environment (Clemmensen and Abrahamsen, 1983). The resistant laminae interbedded between the light-toned, wind-ripple sets appear analogous to “horizontal alternating beds” associated with alternating wet and dry conditions in modern and ancient sand sheet deposits, where fine sand from suspension adhere to a slightly damp truncation surface before being overlain by tractional wind-ripple sets (Lea and Waythomas, 1990; Murton et al., 2025). Alternate explanations such as deflation to the water table accompanied by brief deposition from surface ponding could also be responsible (Chakraborty and Chakraborty, 2001; Clemmensen and Abrahamsen, 1983). Regardless, early diagenesis would preferentially exploit the subtle textural differences between the resistant laminae and the wind-ripple deposits - in a similar fashion to cementation of pinstripe laminae (e.g. Fryberger and Schenk, 1988), causing both the dark-toned appearance of the lamina and the induration. These resistant laminae were likely made more pronounced through wind etching (Fryberger and Schenk, 1988; Murton et al., 2025) creating deep shadows and perhaps acting as traps to accumulate dark basaltic sand, leading to the dark-toned appearance from farfield images and orbit.

3.5.1.5 Do the alternating light- to dark-toned planar bedded lithofacies record fluctuations in the climate?

The repetitive transitions from light to dark-toned planar bedded lithofacies, aeolian scours, and back to light-toned planar bedded lithofacies may have arisen from more long-term changes in the balance between relative water-table level, surface cementation, wind energy, and sediment supply. These factors could have been impacted by changes in the regional climate (Figure 3.20; Mountney, 2012). A warmer, humid climate would lead to water table rise and result in damp surface conditions. A more arid climate would lead to water table level fall and result in either dry surface conditions or deflation depending on the sedimentation rate. The gradational change from the light-toned to dark-toned lithofacies in the Marker Band valley succession may suggest a gradually humidifying and warming climate to explain the decrease in sediment input relative to the rate of water table rise over time. The overlying broad aeolian scours which truncate the dark-toned planar bedded lithofacies could indicate a more arid climate and a fall in the relative water table, triggering intense erosion and the local deflation of partially cemented sandsheets into blowout structures (Roberts et al., 2025). A large increase in sediment availability, compared to the relative water table, would lead to sustained deposition of the dry sandsheets (Figure 3.20). This same gradational change from light- to dark-toned, planar-bedded lithofacies followed by subsequent scouring was also noted in sandsheet deposits on Mars by Grotzinger et al. (2005) in Meridiani Planum and also attributed to a regional climate potentially influencing the relative water table level.

Alternatively, the coeval presence of blowouts, isolated dunes, and both dry and damp sandsheets in terrestrial desert environments (Mountney, 2006b) raises the possibility that the observed light and dark-toned, planar bedded lithofacies could also be attributed to the impact of internal processes affecting the balance between wet and dry deposition and erosion over space rather than time (Figure 3.20; Mountney, 2012). Sediment transport and deposition depends on whether wind is over- or under-saturated with sediment (e.g. Kocurek, 1996). As a wind moves across areas with sediment trapped by a high-water table or surface cementation, sediment mobilisation thresholds are increased, potentially causing winds to become undersaturated. Then, as this wind encounters drier, or less-cemented zones, localised erosion may occur in the form of blowouts (Hesp, 2002; Smyth et al., 2019). Deflation and entrainment of sediment within the blowout could lead to a modest rise in sediment supply. The saturated wind - accompanied by flow deceleration over blowout rims (e.g. Smyth et al., 2019) - would lead to deposition and the accumulation of dry, aeolian sandsheets. In this hypothesis, co-existing dry to damp to erosional desert environments may have accumulated as this sand sea migrated, without needing a changing climate to explain vertical transitions. However, further work on integrated sandsheet and blowout fields is needed to assess how spatial heterogeneity in these systems can influence wind saturation and sediment transport.

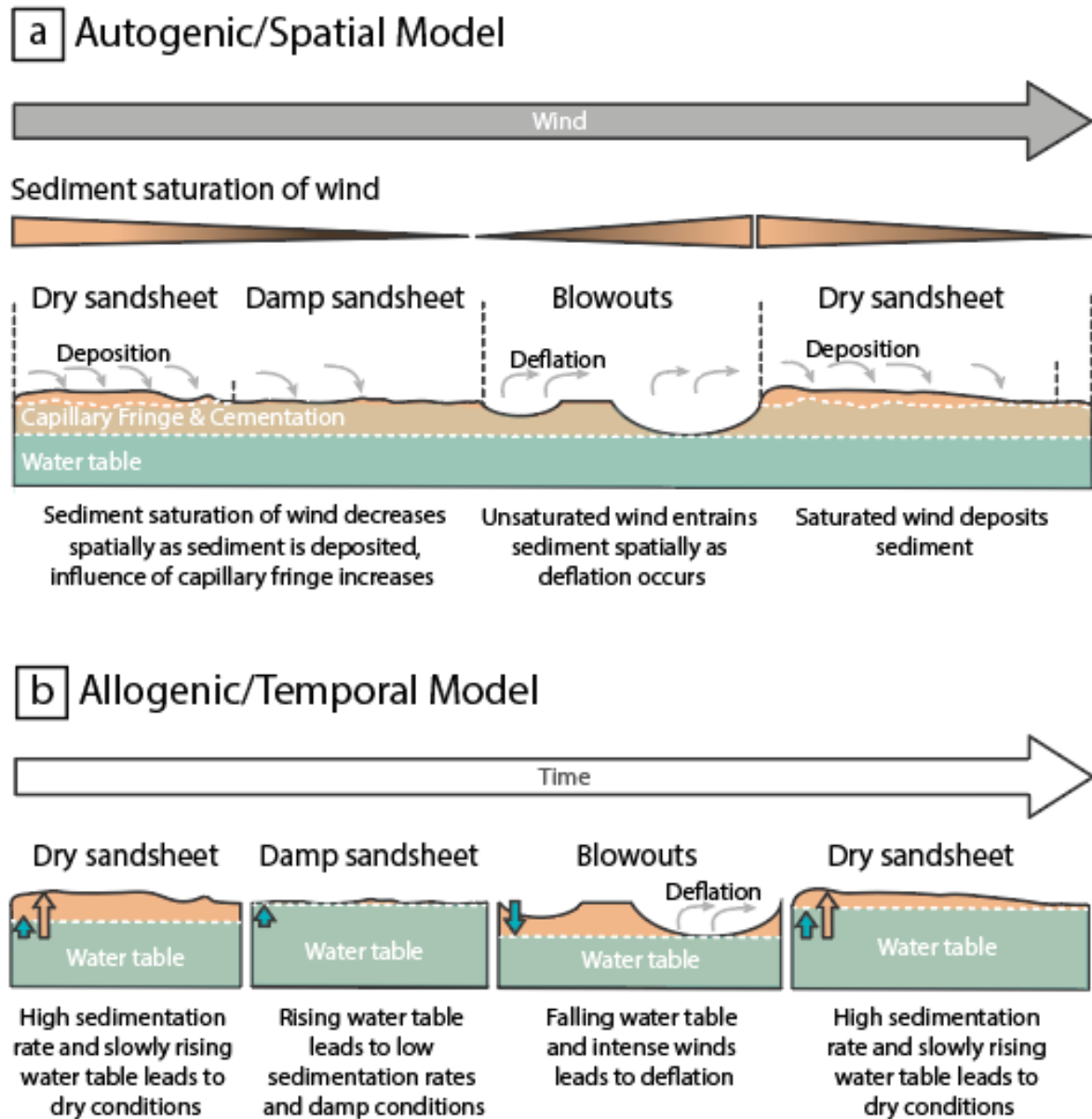


Figure 3.20: Proposed models showing how the occurrence of dry sandsheets, damp sandsheets, and aeolian deflation can occur through internal system dynamics or external factors. (a) Model showing the spatial complexity of co-existing environments in a sand sea impacting the sediment saturation of a given wind which in turn impacts the depositional environment. (b) Alternate dynamic model showing how a climate-controlled relative water table height can impact depositional environments over time.

3.5.1.6 Stratigraphic synthesis and a record of a changing climate at the base of the layered sulfates

From combined sedimentary facies and architectural-element analysis and the above considerations, I construct the following scenario as a possible sequence of events in the depositional history of the basal succession of the layered sulfate-bearing strata (Figure 3.21):

1) Deposition in the orbitally defined Layered Sulfate unit began with the accumulation of simple aeolian dunes migrating over a potential, regionally extensive supersurface in the upper section of the Contigo member (see Section 5.1), indicative of a dry aeolian system with high sediment availability. Minimal influence of the water-table may have generated the flat accumulation surfaces (e.g. Kocurek and Day, 2018; Kocurek and Havholm, 1993; Veiga et al., 2002).

2) A reduction in sedimentation rate relative to a gradual rise in water table led to the inhibition of dune development and promoted the formation of laterally continuous sandsheets.

3) Further decline in sediment input resulted in a depositional hiatus, enabling deflation down to the capillary fringe of the water table and development of a regional deflationary supersurface (Mondro et al., 2025b; Rapin et al., 2021). This surface was modified by surface alteration (Farrand et al., 2025) and contraction-expansion features attributed to thermal contraction of partially cemented sand (e.g. Kocurek and Hunter, 1986; Simpson and Loope, 1985).

4) A modest rise in the water table facilitated the formation of a shallow, regionally extensive lake under ice-free conditions, leading to low-energy subaqueous sedimentation in the Amapari member (Mondro et al., 2025a; Mondro et al., 2025b).

5) Subsequent lake-level fall exposed the subaqueous deposits to aeolian activity. Undersaturated winds reworked or removed these fine-grained lacustrine sediments (e.g. Clemmensen and Tirsgaard, 1990), leading to inferred deflation prior to the accumulation of the Chenapau member.

6) A rise in sediment flux reinitiated subaerial deposition. Surface water was still present as localised runoff in small, fluvial streams as seen by the lens in unit E of the Chenapau member.

7) A net rise in the water table – at the same pace as sediment deposition, led to the accumulation of the thick succession of sandsheets. Short-term climatic variability or autogenic processes drove repeated transitions between damp sandsheet formation, localised deflation, and dry sandsheet accumulation. These cycles reflect fluctuations in the balance between the water table, sedimentation rate, and wind strength.

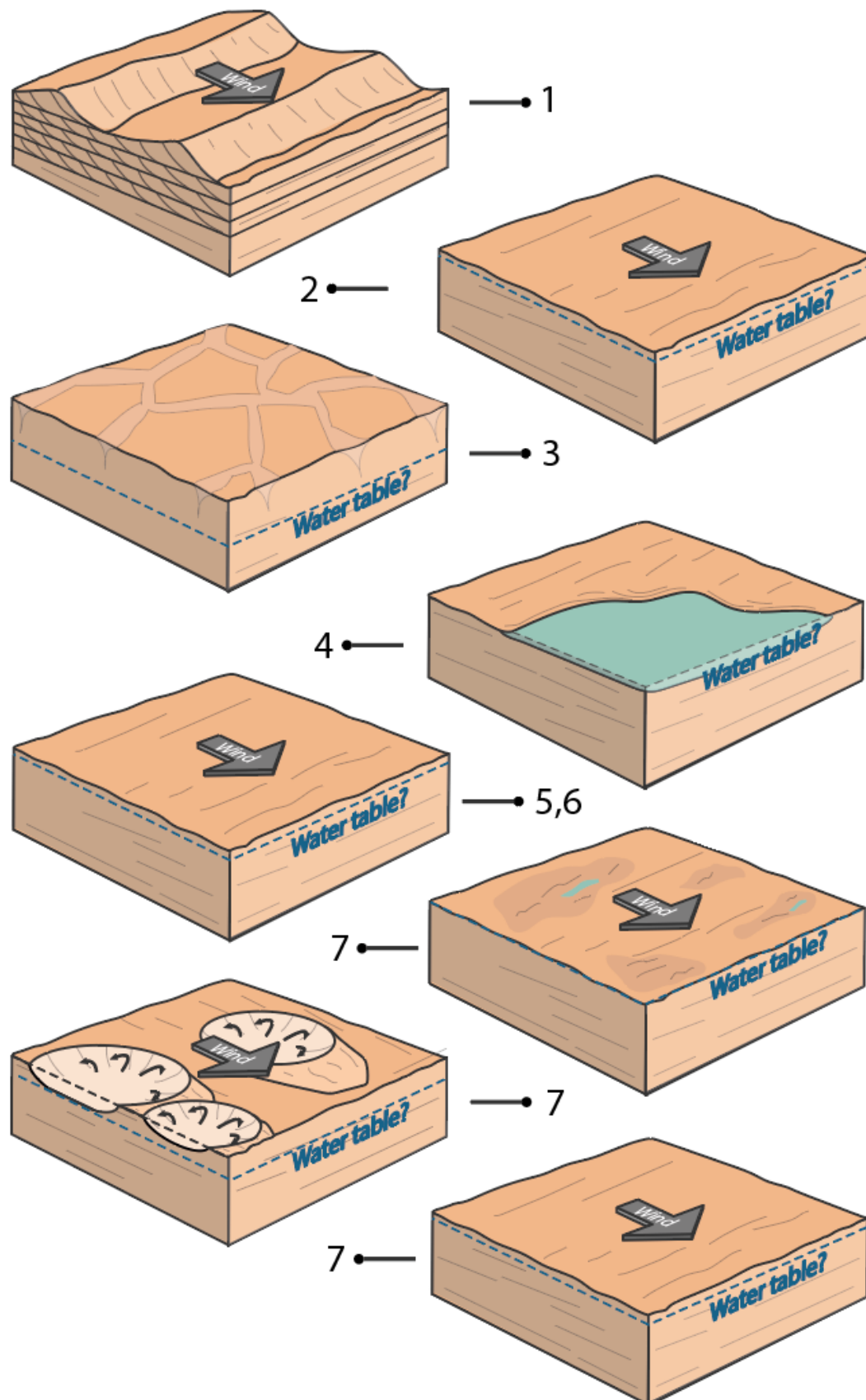


Figure 3.21: Dynamic model showing the stratigraphic synthesis and evolution of facies in Marker Band valley recorded in the Contigo (1), Catrimani (2-3), Amapari (4), and Chenapau (5-7) members as described in Section 5.1.6.

5.5.2 Implications

The orbital-scale transition from clay-bearing strata to a mixed clay- and sulfate-bearing unit, and to the layered sulfate-bearing unit within the stratigraphy of Aeolis Mons (e.g. Fraeman et al., 2016; Sheppard et al., 2021b), has been widely interpreted as a record of a gradual aridification of Gale crater (e.g. Milliken et al., 2010) during the Hesperian, 200 – 500 Myr after Gale crater formed (Grant et al., 2014; Grotzinger et al., 2015; Thomson et al., 2011). Similarly, the transition from lacustrine (Grotzinger et al., 2015; Grotzinger et al., 2014) to shoreline (Gwizd et al., 2024) environments recorded in clay-bearing strata, and finally to aeolian dune environments punctuated by brief episodes of flooding (Edgar et al., 2024; Gupta et al., 2023b) recorded in the mixed clay- and sulfate-bearing unit, implies a climatic shift from wetter to drier conditions. However, our depositional model for the layered sulfates challenges this monotonic aridification trend, also supported by earlier long-distance observations indicated shifting depositional environments in the LSu (Rapin et al., 2021).

The transition from aeolian, simple dunes, to the presence of a thick, laterally extensive succession of aeolian sandsheets indicates that, rather than a simple decline in water table due to continuous evaporation, conditions likely involved an overall *rise* in the water table relative to the depositional surface (e.g. Kocurek and Havholm, 1993; Mountney and Thompson, 2002). On Earth, a relative rise in water table results from subsidence of the sediment column through a fixed water table or an absolute increase in groundwater level (Mountney, 2006a). As Gale crater lacks a mechanism for sustained tectonic subsidence (Grotzinger et al., 2015), the most likely scenario for aeolian sandsheet accumulation and preservation is an absolute - but slowly and maybe intermittently - rising water table that kept pace with the accumulation of the aeolian sandsheets. This leads to the question: could the accumulation of the Mount Sharp group represent relative regional climate warming to account for a slowly rising water table, and was the aridification instead caused by high sedimentation rates infilling the basin that outpaced and overtook water table rise? Infilling of Gale basin under stable climate regime has been proposed as a plausible origin of drying of depositional environments (Horvath and Andrews-Hanna, 2021) but likely, further temporal complexities to the regional climate may have also caused local oscillations in the level of water table, potentially impacted by seasonal melting (e.g. Kite et al., 2013a) or internal dynamics of the aeolian sandsheet system. Such fluctuations caused a cyclical stacking pattern of dry interdune deposits, damp sandsheet conditions caused by intermittent water table rises to the surface, followed by evaporation and near-surface sulfate cementation as chemical data suggests for underlying sulfate enrichments (e.g. Rapin et al., 2023a; Rapin et al., 2019), and aeolian deflation during periods of water table level fall. While the layered sulfates do record a more arid palaeoenvironment than the lacustrine mudstones of the clay-bearing Murray formation, they were also accompanied by a slowly rising groundwater table and intermittent surface water.

Geomorphic evidence and modelling indicate that Gale once had a larger groundwater system extending across northern Aeolis (Putnam and Palucis, 2021). The depth at which the Marker Band

valley succession accumulated – several kilometres below the regional topography – suggests near-surface groundwater may have been restricted to topographic lows, with most of the surrounding terrain arid. However, these topographic lows – with persistent ground water and intermittent surface water – may have acted as localised, sustained habitable niches, with favourable environments for primitive lifeforms to persist (e.g. Westall et al., 2013) and even for prebiotic evolution (Clark et al., 2021; Rapin et al., 2023b).

3.6 Conclusions

This chapter provides a detailed stratigraphic analysis of the ~100 m thick sedimentary succession of the lower layered sulfate-bearing unit comprising the Contigo, Catrimani, Amapari, and Chenapau members of the Mirador formation. The succession is predominantly deposited by aeolian processes dominated by m-scale, tabular cross-bedding in the Contigo member interpreted as aeolian dunes and a light- and dark-toned planar-bedded lithofacies in the Catrimani and Chenapau members predominantly comprised of pinstripe laminae interpreted as aeolian sandsheets. A high-water table likely promoted a flat depositional surface ideal for the accumulation of highly planar and parallel wind-ripple laminae. Episodic surface water occurring as a shallow lake in the Amapari member and a small fluvial lens embedded in the Chenapau member further support a high-water table during the accumulation of the lower layered sulfates. Variations between the light- and dark-toned planar-bedded lithofacies in the Chenapau member reflect differences in sediment availability, moisture content, wind energy, and depositional rate. Darker units likely record wetter conditions with intermittent deposition and potential adhesion structures, while lighter units reflect drier, more continuous aeolian deposition. The ~60 m thickness of aeolian sandsheets far exceeds typical terrestrial analogues. This thickness likely resulted from a slowly rising water table that limited sediment mobilisation. Rather than a steady drying trend, the layered sulfate-bearing unit of Gale crater records intervals of rising water table and persistent subsurface water. While surface conditions were still overall drier than during deposition of the clay-bearing stratigraphy, the persistence of groundwater suggests that habitable niches may have endured longer than a straightforward aridification model would imply.

Chapter 4: Reconstructing Depositional Environments of the Sediment Fan in Oxia Planum, Mars

4.1 Introduction

Fan-shaped sedimentary deposits are easily recognized from their planform geomorphic expression on the surface of Mars and offer valuable insights into the planet's history of surface water, climate evolution, and past habitability (Edgett and Sarkar, 2021; Irwin et al., 2005; Malin and Edgett, 2000; Moore and Howard, 2005). These landforms typically form where a sediment transport capacity of a river declines, typically due to flow deceleration, causing rapid deposition and/or channel avulsion (e.g. Leenman and Eaton, 2021; Moscariello, 2018). Depending on the prevailing environment, sediment can be deposited as a delta (Kleinhans, 2005), fluvial fan (Weissmann et al., 2013), alluvial fan (Blair and McPherson, 1994), or a submarine fan (Maier et al., 2020), all of which have been proposed for Mars's geologic record (e.g. Fawdon et al., 2018; Metz et al., 2009a; Wilson et al., 2021). Some Martian sediment fans may have been active for tens to hundreds of thousands of years (Bhattacharya et al., 2005; Irwin et al., 2015; Lapôtre and Ielpi, 2020), suggesting that fluvial systems, and the climatic conditions supporting them, persisted far longer than short-lived runoff events. Putative deltas have also been used as indicators for ancient standing bodies of water (Fawdon et al., 2018; Mangold et al., 2021; Tebolt and Goudge, 2021), however, without recognisable structures such as clinoforms (e.g. Mangold et al., 2021) distinguishing deltas from fluvial fans from orbit can be challenging. Understanding how these fans formed, and how long they were active, is therefore key to reconstructing surface hydrology, depositional settings, and regional climate evolution. Sediment fans are also high-value targets for robotic exploration, due to their potential to preserve biosignatures and evidence of habitable conditions (Goudge et al., 2017; Mangold et al., 2021; Quantin-Nataf et al., 2021; Summons et al., 2011).

A ~14 km long sediment fan is located at Oxia Planum, within the eastern margin of the landing ellipse for the European Space Agency (ESA) *Rosalind Franklin* rover mission. Along with the sediment fan, a wide-spread phyllosilicate-bearing unit and abundant sinuous ridges show an extensive history of extensive deposition and modification by surface water in the area (Bowen et al., 2022; Davis et al., 2023; Fawdon et al., 2022; Mandon et al., 2021; Quantin-Nataf et al., 2021). The sediment fan is located at the terminus of a ~250 km long valley system, Coogoon Vallis, covering a catchment area of $\sim 2.1 \times 10^5 \text{ km}^2$ - roughly equivalent to Great Britain - which records evidence for multiple major episodes of surface run-off (Fawdon et al., 2022). As the main objective for *Rosalind Franklin* is to search for evidence of past life on Mars (Vago et al., 2017), understanding the depositional processes that formed the sediment fan, is critical for reconstructing the past habitability of the region. Previous studies have conducted broad-scale investigations of this fan, proposing both fluvial

fan and delta depositional models (Gary-Bicas and Rogers, 2021; Ivanov et al., 2020; Quantin-Nataf et al., 2021). To better constrain the origin and depositional history of the Oxia Planum sediment fan, detailed photogeological analysis of its internal architecture, stratigraphic relationships, and surface morphology is required.

The aim of this chapter is to reconstruct the depositional environments and processes that formed the Oxia Planum sediment fan. The main objectives are to: (1) describe and characterize the photogeologic units and geomorphic features associated with the sediment fan; (2) interpret the depositional processes and environments that formed the planform architecture; and (3) discuss the implications for the formation of the sediment fan in Oxia Planum and their broader, regional significance.

4.2 Context

Oxia Planum is located at the transition between Noachian highlands, Arabia Terra, and Hesperian/Amazonian lowlands, Chryse Planitia where a Noachian/Hesperian valley-lake system, Coogoon Vallis, flowed over this area from Arabia Terra to Chryse Planitia (Fawdon et al., 2018) (Figure 4.1). Oxia Planum is a region of scientific interest as a selected rover site as it contains Noachian basement material (Pan et al., 2019), bears a relatively flat topography (Fawdon et al., 2021), a long-lived sedimentary depositional system recorded through layered sedimentary rocks (McNeil et al., 2021), and evidence of sustained aqueous activity through altered phyllosilicate-bearing Noachian deposits (Bowen et al., 2022; Mandon et al., 2021; Quantin-Nataf et al., 2021). Oxia Planum is of further scientific interest as it records evidence for alteration from volcanism (Michalski and Bleacher, 2013), tectonism (Anderson et al., 2001; Woodley et al., 2021), aeolian processes (Favaro et al., 2021) and impact cratering (Roberts et al., 2021).

The sediment fan is located at the eastern margin of the Oxia Planum landing ellipse in an area informally named Raetia Palus (Fawdon et al., 2021) (Figure 4.1). Previously, orbital mapping attempts of this area have been focused on regional geology or the general geology of the landing ellipses. The United States Geological Survey (USGS) recorded a coarse-scale boundary between the two highland units, mNh and INh (Tanaka et al., 2014) at the site of the sediment fan. Since then, more in-depth mapping of the area has been conducted. The fan was first identified by Quantin-Nataf et al. (2016) using OMEGA data (at 2500 m/pix) and CRISM multispectral data (at 200 m/pix and up to 20 m/pixel in certain areas) who found that a unit distinct from the basement rocks exhibited fan-like morphology.

This fan-like unit has since been interpreted to be fluvial in origin (e.g. Fawdon et al., 2022). This fluvial interpretation was due to the lack of mafic signatures, putative cross-bedding (Gary-Bicas and Rogers, 2021), and evidence of aqueous activity through the findings of layers with hydrated silica, Al-phyllosilicates, and Fe²⁺ smectites (e.g. Carter et al., 2016; Carter et al., 2023; Loizeau et al., 2019). Previous studies have identified multiple episodes of deposition (Ivanov et al., 2020; Quantin-Nataf et al., 2021) through multiple craters with diameters between 250 – 320 m which were interbedded

between layers of the sedimentary fan.

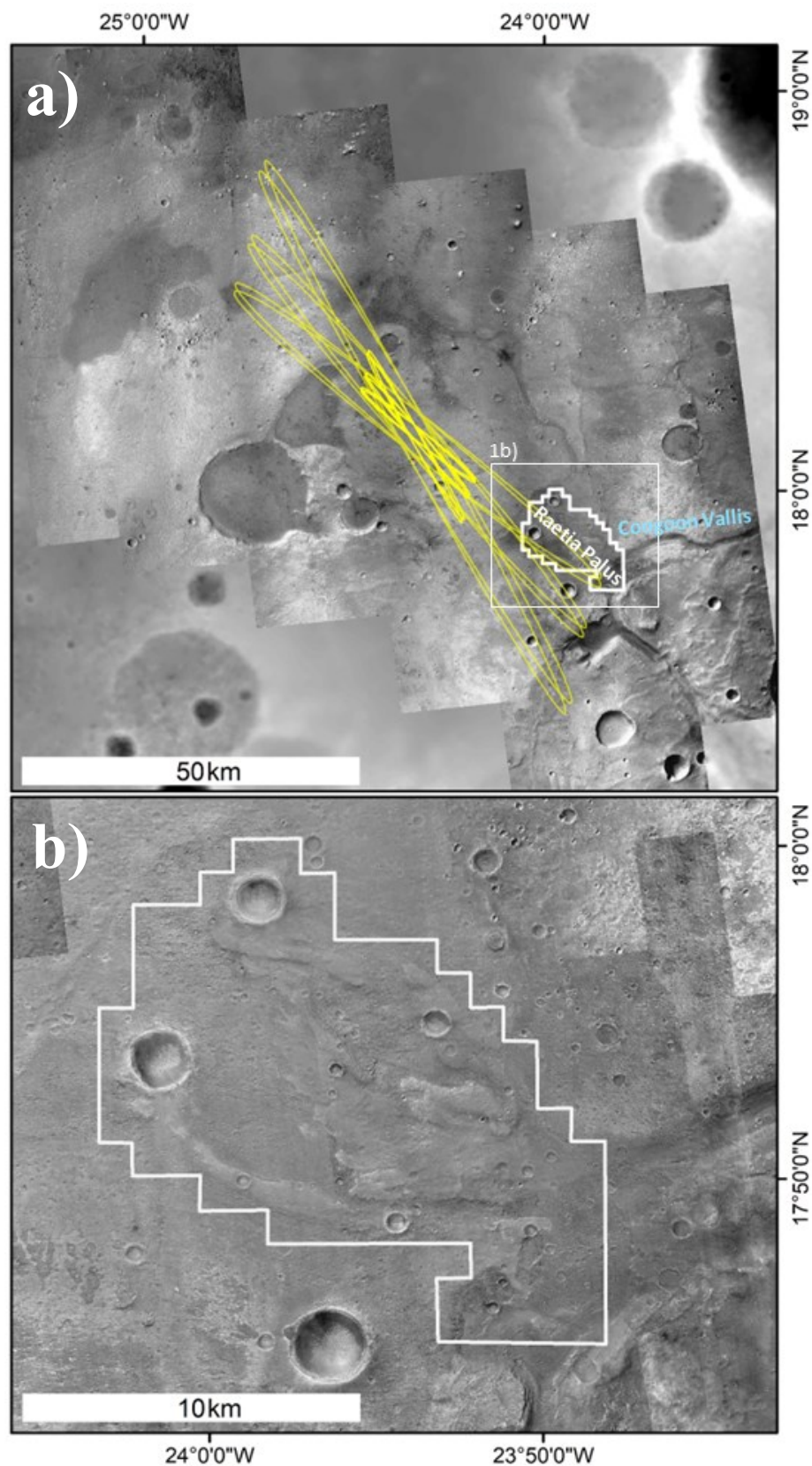


Figure 4.1: a) CTX mosaic (Fawdon et al., 2021) overlaying the HRSC DTM (Ferguson et al., 2018) of Oxia Planum with the Rosalind Franklin rover's 2028 landing ellipses (yellow) and the mapping extent (in white). b) HiRISE (McEwen et al., 2007) mosaic of the sediment fan of Oxia Planum mapping extent (in white).

4.3 Methodology

4.3.1 Datasets

Multiple regional datasets were used to distinguish between tone and morphology in this mapping project. An orthorectified mosaic (Fawdon et al., 2021) of Oxia Planum from the Context Camera was used as a basemap for the regional context (Figure 4.1a). The Context Camera (CTX; Malin et al., 2007) consists of a single panchromatic band, broadly covering 500-700 nm wavelengths, with image resolutions of ~6 m/pixel, enabling recognition of broad changes in texture and tone in bedrock. False colour images consisting of a combination of the near-infrared, panchromatic, and blue-green bands (NPB; covering 850-475 nm wavelengths) from the Colour and Stereo Surface Imaging System (CaSSIS; Thomas et al., 2017) at 4.5 m/pixel resolutions were also used to distinguish textural and tonal differences between bedrock.

To divide the bedrock into distinct units, map their extents, and identify additional geomorphic and architectural features, I created a high-resolution mosaic using images from the High-Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007), orthorectified to the CTX basemap of Oxia Planum (Figure 4.1b). This mosaic has a resolution of 25-50 cm/pixel and is constructed from HiRISE RED band images, broadly sensitive to 550-850 nm wavelengths. This mosaic enabled detailed investigations into metre-scale textures of bedrock, geomorphology, and broadscale architecture of the sediment fan.

Regionally, a CTX Digital Terrain Model (DTM) mosaic with a horizontal post-spacing of 20 m/pixel (Fawdon et al., 2021) was used to identify the topography of the mapping area. This CTX DTM mosaic was locally augmented with HiRISE DTMs I created (see Appendix III.1) from stereopairs calibrated to the MGS Mars Orbiter Laser Altimeter. (MOLA; Zuber et al., 1992). The HiRISE DTMs have a horizontal post-spacing of 1-2 m/pixel and cover most of the surface area of the sediment fan.

4.3.2 Mapping and Identification of Units

Stratigraphic units in the sediment fan and surrounding area were identified and mapped based on principles of photographic recognition of sedimentary rocks developed for Mars (Edgett and Sarkar, 2021). According to these principles, geologic attributes such as relative albedo, colour, texture, pattern, erosional expression, crater retention, and topography were used to distinguish between mapped units similar to previous detailed photogeologic maps made of sediment fans on Mars (Anderson and Bell III, 2010; Rice et al., 2013; Stack et al., 2020b). In order to ensure methodological consistency, a 1x1 km grid was generated. 150 individual 1x1 km grid cells were mapped at a scale of 1:1000 and subjected to double verification.

4.3.3 Unit Descriptions and Interpretations

Within and surrounding the sediment fan at Oxia Planum, a total of nine units with different albedos, photo-interpreted textures, and elevations were identified as bedrock for their boulder production, preservation of steep slopes, and impact crater retention (Edgett and Sarkar, 2021) (Figure 4.2). Four units occur stratigraphically below the sediment fan and comprise the basement of Raetia Palus: a raised linear feature bearing unit (RL), a light-toned, fractured unit (LtF), a dark-toned, pitted unit (P), and a sub-polygonally fractured unit (SF). Four distinct bedrock units were recognised within the sediment fan system: an upper fractured unit (UF), a ledge-forming unit (LF), a Medium-light Toned Smooth unit (MltS), and a Medium-light Toned Rough unit (MltR). A Light-toned crater fill unit (LtCF) was also identified within circular topographic lows on the sediment fan and the surrounding area.

Here, the photo-interpreted textures, tones, and stratigraphic relationships are described and depositional processes are interpreted for each unit in stratigraphic order. These units are summarised in table 4.1 and shown in Figure 4.2 along with crater ejecta and crater basin units.

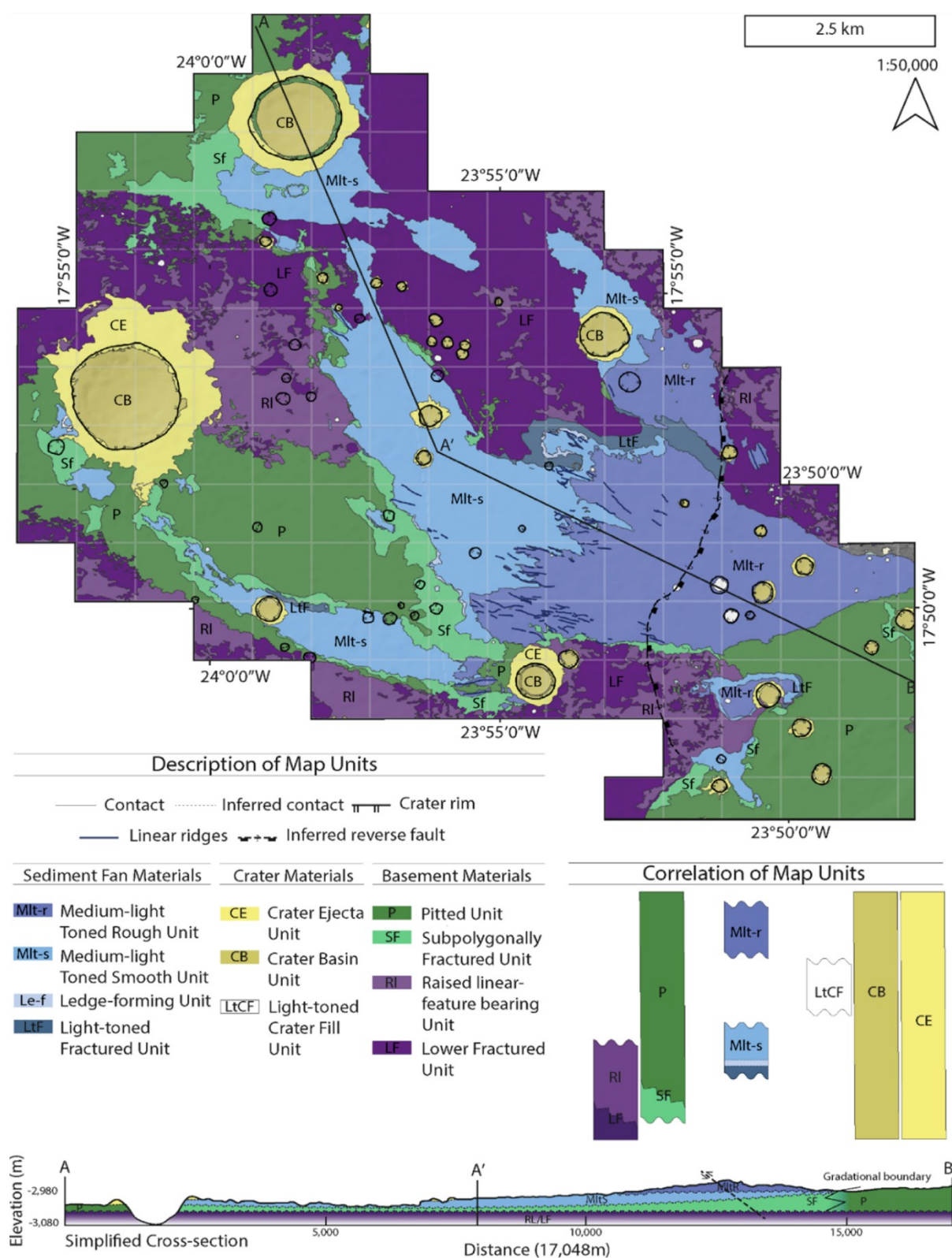


Figure 4.2: Photogeologic map of inferred stratigraphic units of the Oxia Planum Sediment Fan and associated units, cross-section with a 6 x vertical exaggeration, and correlation of map units, mapped by 1x1 km squared grid (light-grey).

Unit	Unit name - Description HiRISE & CaSSIS (Near Pan and Blue; NPB)	Spatial Outcrop Extent	Elevation Range (m)	Additional information (Unit associations, other details, compositional)	Interpretations
RI	Raised Linear-Feature Bearing unit - HiRISE; Mottled medium-to dark-toned, raised linear features forming polygons. CaSSIS NPB; bright and medium-to dark-toned blue hues.	~9.3 km ²	-3042 to -2962	Gradational contact with Light-toned Fractured unit.	Polygonal ridges likely formed by diagenetic cementation of fracture networks; possibly linked to fluid movement.
LF	Lower Fractured unit - HiRISE; Mottled light-toned, fractured. CaSSIS NPB; light-toned blue hues.	~28.3 km ²	-3037 to -2950	Gradational contact with unit RI. Individual fractures can be traced into raised linear features within unit RI.	Phyllosilicates formed either from deposition or alteration from aqueous processes. Fracture network the same as the one that caused the raised linear features in unit RI.
P	Pitted unit - HiRISE; Dark-toned, m-scale rough texture caused by a pock-marked, pitted morphology, steep scarps, pristine impact craters, boulder shedding. CaSSIS NPB; dark-blue with orange-yellow highlights	~32.4 km ²	-3047 to -2925	Forms bench-shaped contacts against underlying LF and RI; gradational contact with Sf.	Potential dust or ash fall (Harris et al., 2024a). Roughness caused by erosion.
Sf	Sub-polygonal Fractured unit - HiRISE; Medium-toned, sub-polygonal fractures, smoother and pockmark-free compared with adjacent unit P. CaSSIS NPB; dark-blue.	~9.25 km ²	-3041 to -2942	Gradational contact with P; rare bench-shaped contacts with underlying LF and RI.	Less degraded version of the Pitted unit; potentially protected from erosion by sediment fan.
LtF	Light-toned Fractured unit - HiRISE; Light-toned, deep curvilinear fractures forming circular “puzzle-piece” blocks. CaSSIS NPB; bright white- to light blue-tones.	~1.37 km ²	-3037 to -2950	Underlies Mlt-s and Ledge-forming unit.	Possible early fan-related deposit.

Le-f	Ledge-forming unit – HiRISE; Medium-light toned fractured ledge unit, thin (~1 m), found at fan scarp base. CaSSIS NPB: bright white- to light blue-tones.	~0.11 km ²	-3030 to -3026	Bench-shaped contacts observed with underlying U-F and Sf.	Possible early fan-related deposit.
Mlt-s	Medium-light Toned Smooth unit – HiRISE; Medium-light tone, smooth texture, planar and inclined layering, polygonal fracturing. CaSSIS NPB; medium-blue tones.	~22.9 km ²	-3038 to -2903	Overlies U-F and Ledge-forming unit; underlies Mlt-b.	Prolonged deposition in a deltaic or fluvial fan environment.
Mlt-b	Medium-light Toned Rough unit – HiRISE; Medium-light toned, rough metre-scale protrusions, boulders, curvilinear fractures, striations. CaSSIS NPB; medium-blue tones.	~17.6 km ²	-3006 to -2946	Overlies Mlt-s, infills craters, contains Light-toned Crater Infill unit.	Short-lived fluvial fan deposit with inverted fluvial channels, possibly coarse-grained which may indicate flooding.
LtCF	Light-toned Crater Infill unit – HiRISE; Light-toned, concentric banding within craters; stands proud; crater-retaining. CaSSIS NPB; bright white- to light-blue tones.	~0.29 km ²	-2966 to -3040	Contained within and partially covered by Mlt-b; found in LF.	Slow evaporation processes or volcanic ash infill.

Table 4.1: Summary of photogeologic units

4.4 Photogeologic Units

4.4.1 Basal units

4.4.1.1 Raised Linear-Feature Bearing Unit (Rl)

Observations

The lowest stratigraphic unit comprising the basin in Raetia Palus is the Raised Linear-Feature Bearing unit (Figure 4.3), outcropping between elevations -3042 m to -2962 m. At a map scale, this unit exhibits a mottled medium- to dark-toned appearance in panchromatic CTX (Figure 4.3a) and HiRISE (Figure 4.3b) images. Tonal variations are also observed in CaSSIS NPB images with predominantly bright and medium- to dark-toned blue hues (Figure 4.3c). This unit is heavily cratered with craters exhibiting an often degraded appearance – often lacking sharp rims and ejecta blankets. While the contact between the Rl unit and the LF is gradational, the contact between unit P, unit SF, and units associated with the sediment fan are all sharp, marked by a distinct bench-shape.

At a m-scale, this unit is characterised by raised linear features 1-2 m wide, with individual ridges reaching up to 150 m. These linear features are often organized into polygonal networks with individual polygons measuring several metres across (Figure 4.3b and d). The polygonal depressions often have a dark-toned appearance while the raised linear features are consistently light-toned in appearance. Occasional exposures of incredibly bright-toned areas within polygons in both HiRISE and CaSSIS images also occur on the $\sim 10^1$ m scale (Figure 4.3b and c). Small, inverted craters - impact craters preserved in positive relief due to resistant fill material - tens of metres in diameter were also identified in this unit and share a similar tone and appearance as the linear features (Figure 4.3d).

Interpretation

As unit Rl underlies all other mapped units, the unit is interpreted to predate the formation of the sediment fan and represent one of the oldest units in the mapping area. The mottled medium- to dark-toned appearance likely reflects the trapping of dark-toned surficial materials within hollows between the network of light-toned raised linear features, obscuring much of the original substrate. The sparse, $\sim 10^1$ m scale bright-toned areas may represent areas where the original bedrock substrate is locally exposed. The distinctive raised linear features are organised into a polygonal network similar to ‘honeycomb’ textures in Oxia Planum (Fawdon et al., 2024), and ‘boxwork structures’ elsewhere on Mars (Anderson and Bell III, 2010; Banham et al., 2024b; Siebach and Grotzinger, 2014). These raised linear features could have formed through differential erosion of filled or cemented fractures post-dating (e.g. Anderson and Bell III, 2010; Siebach and Grotzinger, 2014) or concurrent with (e.g. Banham et al., 2024b; Cardenas et al., 2019) deposition and lithification.

Previous hyperspectral studies have included unit Rl as part of a regionally extensive phyllosilicate-bearing unit (e.g. Ivanov et al., 2020; Mandon et al., 2021; Quantin-Nataf et al., 2021). The

phyllosilicates are attributed to either deposition within a subaqueous environment (authigenic or detrital in origin) or through subsequent aqueous alteration of the surface or near-surface (Mandon et al., 2021; Quantin-Nataf et al., 2021). The bright and medium- to dark-blue tones observable in CaSSIS images, the polygonal network of raised linear features, and relief of unit R1 also resemble the Upper Bedrock Group 2 (Fawdon et al., 2024) identified in HiRISE scale, geomorphic mapping attempts elsewhere in Oxia Planum, may suggest that unit R1 is part of a more regionally extensive unit.

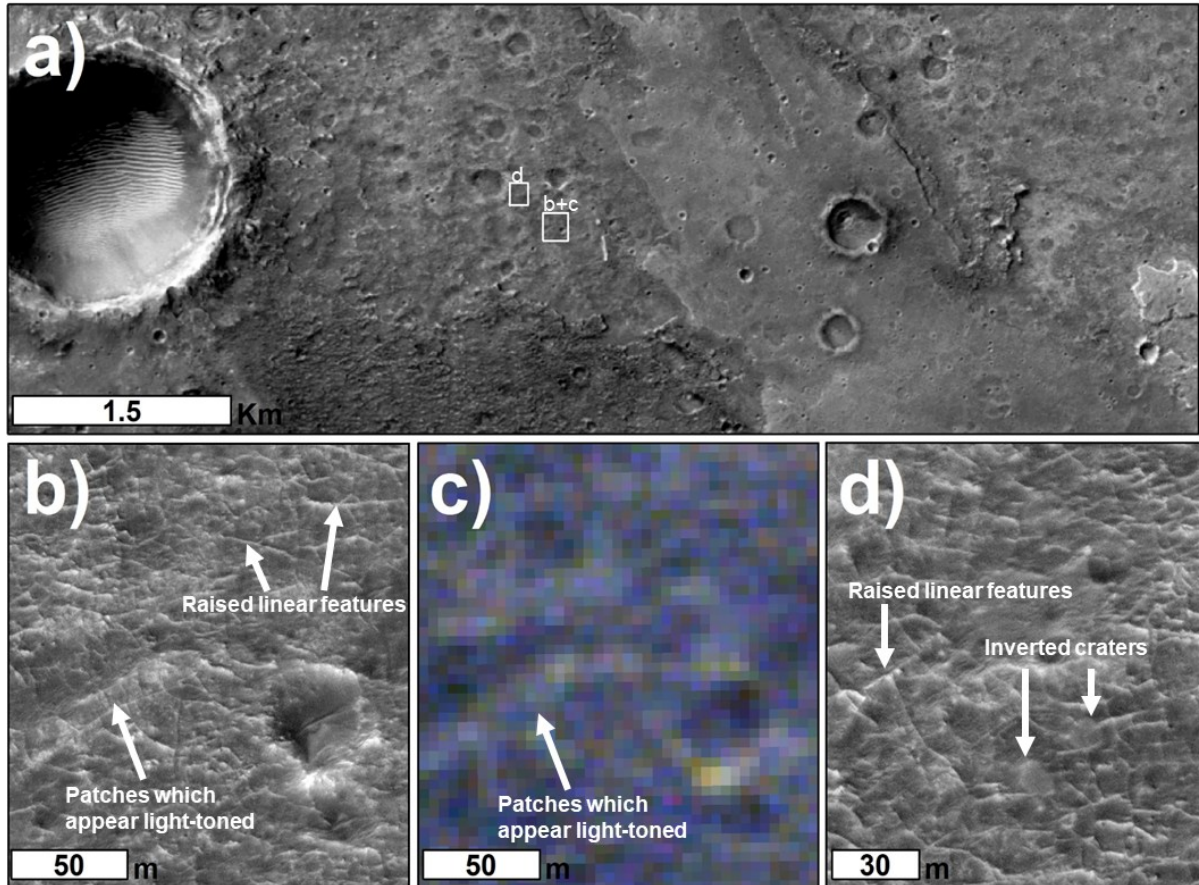


Figure 4.3: The Raised Linear Feature Bearing unit: (a) the context of b, c, and d in the CTX mosaic; (b) the texture and tone of the Raised Linear Feature Bearing unit in HiRISE; (c) the texture and tone of the Raised Linear Feature Bearing unit in CaSSIS; (d) inverted craters similar in tone to the raised linear features within this unit.

4.4.1.2 Lower Fractured Unit (LF)

Observations

The second lowest stratigraphic unit occurring in Raetia Palus is the Lower Fractured unit (Figure 4.4b), outcropping between elevations -3037 m and -2950 m. At a map scale, this unit has a mottled, light-toned appearance in panchromatic CTX (Figure 4.4a) and HiRISE (Figure 4.4b) images, and exhibits light-toned blue hues in the CaSSIS NPB images (Figure 4.4c). Similar to unit Rl, unit LF is also heavily cratered with craters exhibiting an often-degraded appearance, but no inverted craters have been identified within the mapping extent. The contact between unit Rl and LF is gradational (Figure 4.4c) marked by the introduction of raised linear features from unit Rl. The contact between the overlying unit P, unit SF, and units associated with the sediment fan are all sharp, marked by a distinct bench-shape.

At a m-scale, unit LF is fractured and has a mottled light-toned appearance. The fractures cross-cutting this unit are dark-toned in appearance (Figure 4.4b) and can reach up to 150 m lengths and 2 m widths. The fractures interconnect in an extensive network, subdividing the unit into light-toned polygons that are 10 - 50 m in diameter (Figure 4.4b). On some occasions, fractures from unit LF grade into raised linear features from unit Rl (Figure 4.4d).

Interpretations

Similar to unit Rl, unit LF is also interpreted to pre-date the sediment fan. The extensive fracturing of unit LF may be attributed to a variety of processes including impact related, hydraulic related, or contraction and expansion related processes (e.g. Cosgrove et al., 2022). The dark-toned appearance of the fractures is likely caused by dark- to medium-toned surficial material becoming trapped in these crevasses. As multiple fractures in unit LF are observed to grade into raised linear features in unit Rl (Figure 4.4d), the fracturing that caused both the polygonal fracture network in unit LF and the honeycomb texture in unit Rl are likely the same. Post-depositional fluids, therefore, likely migrated through this pre-existing fracture network unevenly, resulting in localised cementing and filling of fractures in unit Rl with erosion resistant material, and may not have as extensively altered or filled fractures contained within unit LF.

Outcrops of the Lower Fractured unit in Raetia Palus have been found to contain hydrated phyllosilicate minerals (Carter et al., 2016) and are included as part of a regional phyllosilicate-bearing unit by other mapping efforts (Bowen et al., 2022; Mandon et al., 2021; Quantin-Nataf et al., 2021). The bright and medium- to dark-blue tones observable in CaSSIS NPB images and the decametre scale extensive fracture networks are consistent with a regionally extensive mineralogic Blue sub-unit (Bowen et al., 2022; Mandon et al., 2021) and related geologic upper Bedrock unit (Fawdon et al., 2024). Whether the phyllosilicates originated from subsequent alteration or were detrital in nature, however, cannot be resolved from the details observed in Raetia Palus.

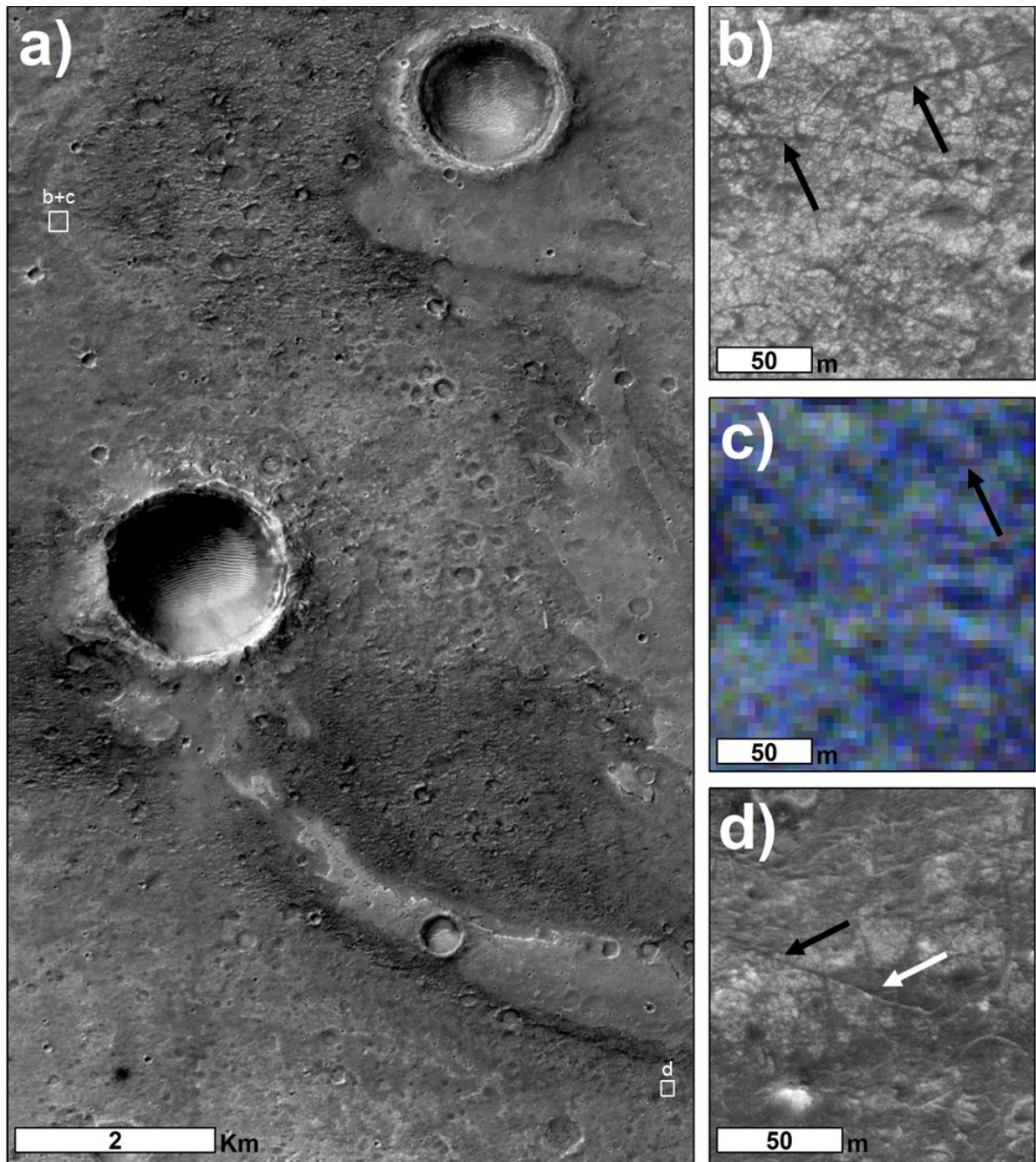


Figure 4.4: The Lower Fractured unit: (a) the context of figure b, c, and d in the CTX mosaic; (b) the texture and tone of the Lower Fractured unit in HiRISE (black arrows show prominent fractures); (c) the texture and tone of the Lower Fractured unit in CaSSIS; (d) arrows show a raised linear feature (white arrow) grade into a fracture (black arrow).

4.4.1.3 Pitted Unit (P)

Observations

The Pitted unit is characterized by a dark-toned appearance at a map scale, recording the lowest albedo in Raetia Palus in both CTX and HiRISE images (Figure 4.5a). In the CaSSIS NPB dataset, this unit exhibits dark-blue tones with orange-yellow highlights (Figure 4.5c). The unit appears to have a rough, pock-marked surface texture resulting from m-scale bumps and ridges (Figure 4.5a). Over the spatial distribution of this unit, there is an apparent degradation of the more continuous Pitted unit into

isolated chains of ridges that extend up to 100 m from the main outcrop. These ridges trend broadly west-northwest/east-southeast (260-280°) (Figure 4.5d). Their appearance and spatial context suggest these ridges are the erosional remnants of the Pitted unit. Here, unit P is in contact with units LR and SF as a sharp, bench-shaped contact with ~1 to 4 m of relief. When unit P is in contact with unit SF, however, there is a gradational boundary marked by a subtle decrease in surface roughness and increase in albedo (Figure 4.5a). Unit P retains a large number of craters and, unlike units LF and RI, the craters often appear fresher, retaining sharp rims and bowl-shapes.

At a m-scale, unit P has a rough texture on the tens of metre-scale comprised of pits and sharp peaks. The peaks exhibit a light-toned appearance while the pits appear more dark-toned (Figure 4.5b). Occasionally, the pits are filled with a surficial material which forms bedforms (Figure 4.5b). This unit is observed to have boulder shedding scarps. The boulders are 1-2 m in size and are light-toned.

Interpretations

Outcrops of the Pitted unit in Raetia Palus have previously been mapped as a regionally extensive Dark Resistant unit (Mandon et al., 2021; Quantin-Nataf et al., 2021) and a Low albedo, Thin Resistant unit (Harris et al., 2024a). Quantin-Nataf et al. (2021) interpreted that the Pitted unit post-dated and onlapped onto the sediment fan, however HiRISE-scale mapping revealed clear bench-shaped contacts between the sediment fan and underlying units SF and P indicating that unit P predated the sediment fan (also see Harris et al., 2024a). Due to mafic spectral signatures found within this unit (Carter et al., 2016), the Pitted unit has previously been hypothesised to be volcanic in origin (Mandon et al., 2021; Quantin-Nataf et al., 2021). However, the observation of pits infilled by a dark-toned, bedform-forming surficial material (Figure 4.5b), likely indicates that the original spectral signatures have been obscured by surficial deposits.

The Pitted unit has a higher competence than units RI and LF as the unit retains a high number of pristine impact craters and the ability to produce and retain boulders (e.g. Edgett and Sarkar, 2021). The apparent degradation of the Pitted unit into isolated chains of ridges along the unit margins appear similar to kharafish or yardangs (e.g. Anderson and Bell III, 2010; Edgett and Sarkar, 2021; Embabi and Embabi, 2018) suggesting significant aeolian wind activity has eroded this unit and likely contributed to the irregular surface texture. The preferred east-southeast and west-northwest trends of the candidate yardangs, are consistent with modern day wind directions observed regionally in Oxia Planum (Favaro et al., 2021).

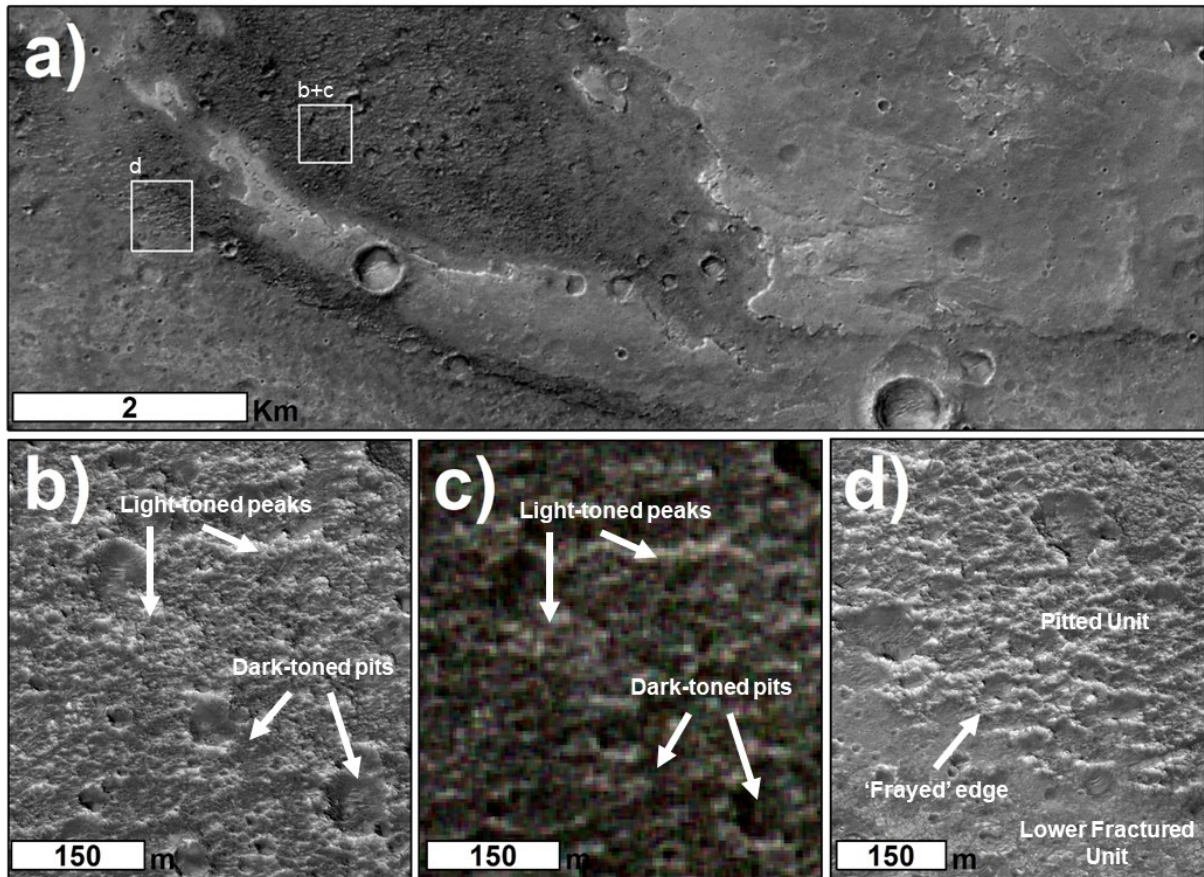


Figure 4.5: The Pitted unit: (a) the context of b, c, and d in the CTX mosaic; (b) the texture and tone of the Pitted unit in HiRISE; (c) the texture and tone of the Pitted unit in CaSSIS; (d) the erosional expression of the Pitted unit.

4.4.1.4 Sub-polygonal Fractured Unit (Sf)

Observations

The Pitted unit transitions into the Sub-polygonal Fractured unit (Sf) – characterized by medium-toned and sub-polygonally fractured bedrock – within ~2 km of the sediment fan margin. The medium-tone is visible in both CTX (Figure 4.6a) and HiRISE (Figure 4.6b) images, and the unit exhibits a darker, bluish hue in CaSSIS NPB images (Figure 4.6c). The fractures appear dark-toned and subdivide the bedrock into 2 – 12 m polygons (Figure 4.6b and d). The surface texture of the unit is relatively smooth and pockmark-free compared to the laterally adjacent unit P. Moving away from the fan and toward the gradational contact with unit P, the Sf unit is transitional in nature and becomes rougher and more irregular in texture, with increased surface pitting and progressively less distinct fractures. In contrast, areas closer to the sediment fan appear smoother and lack visible fractures altogether (Figure 4.6b). While contacts with the underlying units Rl and LtF are rare, these contacts are expressed as sharp, bench-shaped scarps (Figure 4.6d). The contact with the overlying sediment fan units Mlt-S and Mlt-R are also expressed as sharp, cliff-benches. Unit SF is moderately to heavily cratered, with numerous small craters remaining sharply defined (Figure 4.6b).

Interpretation:

Previous research has included outcrops of unit SF with unit P as part of the regionally extensive Dark Resistant unit (Mandon et al., 2021; Quantin-Nataf et al., 2021) and the Low albedo, Thin Resistant unit (Harris et al., 2024a). While this chapter makes a distinction between unit SF and P for their distinct tones and surface textures, the gradational transition from the Pitted unit to the Subpolygonal Fractured unit, coupled with the attributed origin of the Pitted unit's distinct texture to substantial weathering (Section 4.4.1.3), suggests that unit SF represents a less degraded form of the Pitted unit. Unit SF's confinement to areas near the sediment fan implies the fan may have shielded these regions from subsequent erosion.

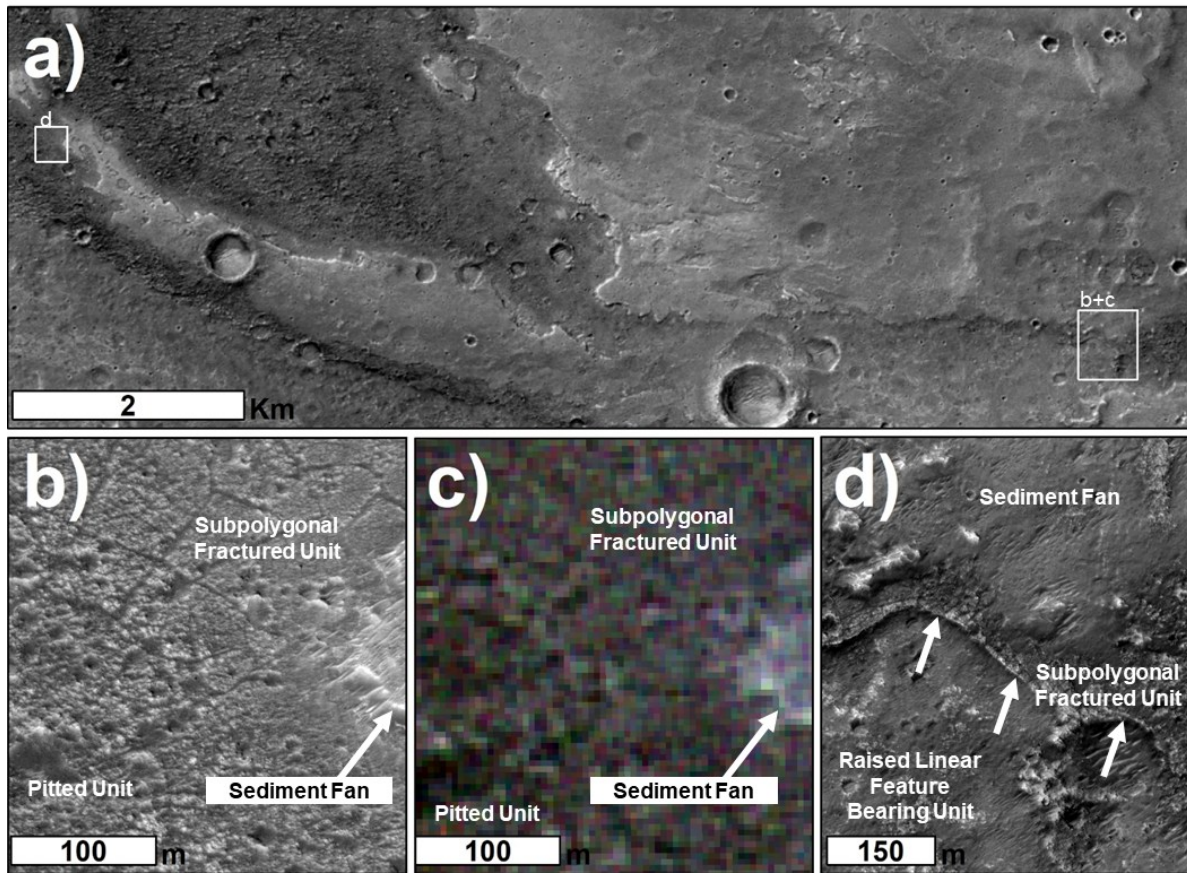


Figure 4.6: The Sub-polygonal Fractured unit: (a) the context of b, c, and d in the CTX mosaic; (b) the texture and tone of the Subpolygonal unit in HiRISE and how it grades into the Pitted unit further away from the sediment fan; (c) the texture and tone of the Subpolygonal Fractured unit in CaSSIS; (d) the erosional expression of the Subpolygonal Fractured unit.

4.4.2 Sediment Fan and Associated Units

4.4.2.1 Light-toned Fractured Unit (LtF)

Observations

The Light-toned Fractured unit (LtF) is the lowest stratigraphic unit directly associated with the sediment fan, occurring between elevations of -3037 m and -2950 m (Figure 4.7a). At a map scale, this unit has a light-toned appearance with the highest albedo recorded in Raetia Palus in both CTX and HiRISE images (Figure 4.7a–b). In CaSSIS NPB images, this unit is distinguishable by a bright white-to blue-toned appearance (Figure 4.7c). High olivine-to-phyllsilicate ratios have been identified here in previous CRISM multispectral analysis (Mandon et al., 2021). The LtF unit occurs primarily within an alcove in the northern portion of the Medium-light Toned Smooth unit (MltS). Only a single ~ 115 m diameter, pristine crater has been identified within this unit.

At a HiRISE scale, the LtF unit is characterised by m-scale, sub-circular blocks. These blocks vary in size, with adjoining blocks fitting together cleanly, like puzzle pieces (Figure 4.7d). Orthogonal fractures are also visible within this unit, continuous for hundreds of metres, and separate the surface of this unit into ~ 50 - 100 m wide sections (Figure 4.7b and c).

Interpretations

The restricted spatial distribution of the Light-toned Fractured unit to within a 1.37 km² alcove of one of the branches of the sediment fan and its absence elsewhere in Raetia Palus likely indicate that unit LtF was either genetically related to and/or protected from degradation by the sediment fan. Unlike units Sf and P that were protected by the sediment fan, unit LtF has a distinct appearance and fracture morphology that differentiate it from any units previously identified in Oxia Planum (e.g. Fawdon et al., 2024), suggesting that this unit is local to Raetia Palus. If this unit is genetically related to the sediment fan and is subsequently a sedimentary rock, the high olivine-to-phyllsilicate ratios identified here (Mandon et al., 2021) may suggest a basaltic sandstone or siltstone composition (e.g. Bedford et al., 2020; Rogers et al., 2018).

The semi-circular blocky texture likely formed in situ through fracturing. While similar patterns can form syndepositionally in sabkha settings due to salt-crust fabrics (Benison, 2019; Goodall et al., 2000), no evaporite minerals have been detected (Carter et al., 2016). Similarly, desiccation cracks in phyllosilicate-rich mudstones are also unlikely due to the high olivine/phyllsilicate ratios (Mandon et al., 2021). More plausibly, the semi-circular blocks formed through polygonal fracturing and turtleback weathering of an isotropic rock or transversely isotropic rock along a bedding plane (Chan et al., 2008; Goehring and Morris, 2005). The unique fracture pattern, tone, and surface texture from the underlying and overlying units suggests that this unit has a unique composition and/or sedimentary texture to the other geomorphic units identified in Raetia Palus.

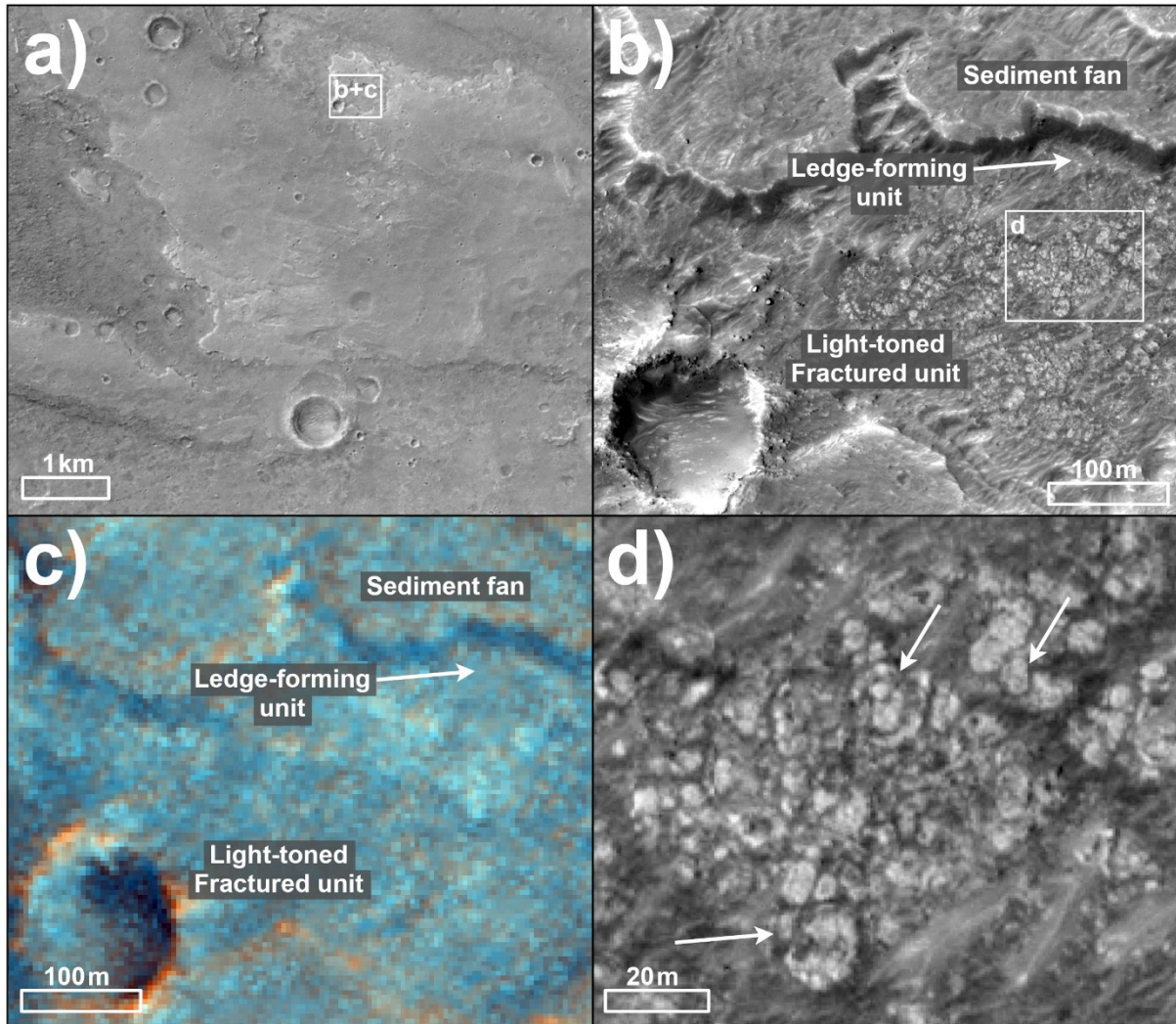


Figure 4.7: The Upper Fractured unit: (a) the context of b and c in the CTX mosaic; (b) the texture and tone of the Upper Fractured unit in HiRISE and its contact with the ledge forming unit; (c) the texture and tone of the Upper Fractured unit in CaSSIS (d) detail of b showing the m-scale sub-circular blocks, arrows show examples of where adjoining blocks fit together like puzzle pieces.

4.4.2.2 Ledge-forming Unit (Le-f)

Observations

The Ledge-forming unit outcrops as narrow ledges (<100 m wide) on top of units LtF and Sf in a few isolated locations. Its exposure is spatially restricted to within ~200 m of the MltS unit. Despite the limited areal extents, unit Le-f can be distinguished from unit LtF for its medium-tone in HiRISE images (Figures 4.7b and 4.8b), and from units Sf and MltS for its distinct bright white- to blue-toned appearance in CaSSIS NPB images (Figures 4.7c and 4.8c). The relative vertical thickness of unit Le-f was measured as ~1 m from the underlying units in HiRISE and CTX DTMs. At one isolated outcrop that is disconnected from the main fan, fractures subdivide the surface into subpolygonal blocks 10-30 m wide (Figure 4.8c). However, when the unit originates out of the fan as a ledge, fractures are not observed.

Interpretations

The spatial restriction of the Ledge-forming unit to within ~200 m of the MltS unit and absence of outcrops elsewhere in Raetia Palus indicate that unit Le-f was genetically linked to and/or shielded from erosion by the overlying fan deposits. The concentration of fracturing primarily in areas farther from the sediment fan supports the interpretation that emplacement of fan-related materials helped protect unit Le-f from extensive degradation. Outcrops of this unit also correspond to multispectral detections of amorphous silica (McNeil et al., 2025a, b; Pan et al., 2021; Turner et al., 2021) that were interpreted as part of a regionally extensive Hydrated Silica-bearing unit extending ~60 km away from the sediment fan (McNeil et al., 2025a, b). A detrital origin for the silica is unlikely, given the absence of a suitable source region within the Coogoon Valles catchment area (Fawdon et al., 2022; McNeil et al., 2025b; Turner et al., 2021). An authigenic source for the opal through evaporation associated with low-water discharge during fan emplacement has also been suggested by (Pan et al., 2021). However, if all hydrated silica detections at Oxia Planum resulted from evaporation associated with fan emplacement, the absence of fan-related units or geomorphic features beyond Raetia Palus would argue against distal deposits related to fan emplacement. Yet, if the sediment fan was a delta, the extent and distribution of silica detections could fall within the proposed bounds of a standing body of water in Oxia Planum (Quantin-Nataf et al., 2021). Alternatively, the opal signatures in unit Le-f could have had a diagenetic origin via silica redistribution through groundwater interaction (McNeil et al., 2025a, b) and postdate the fan's emplacement.

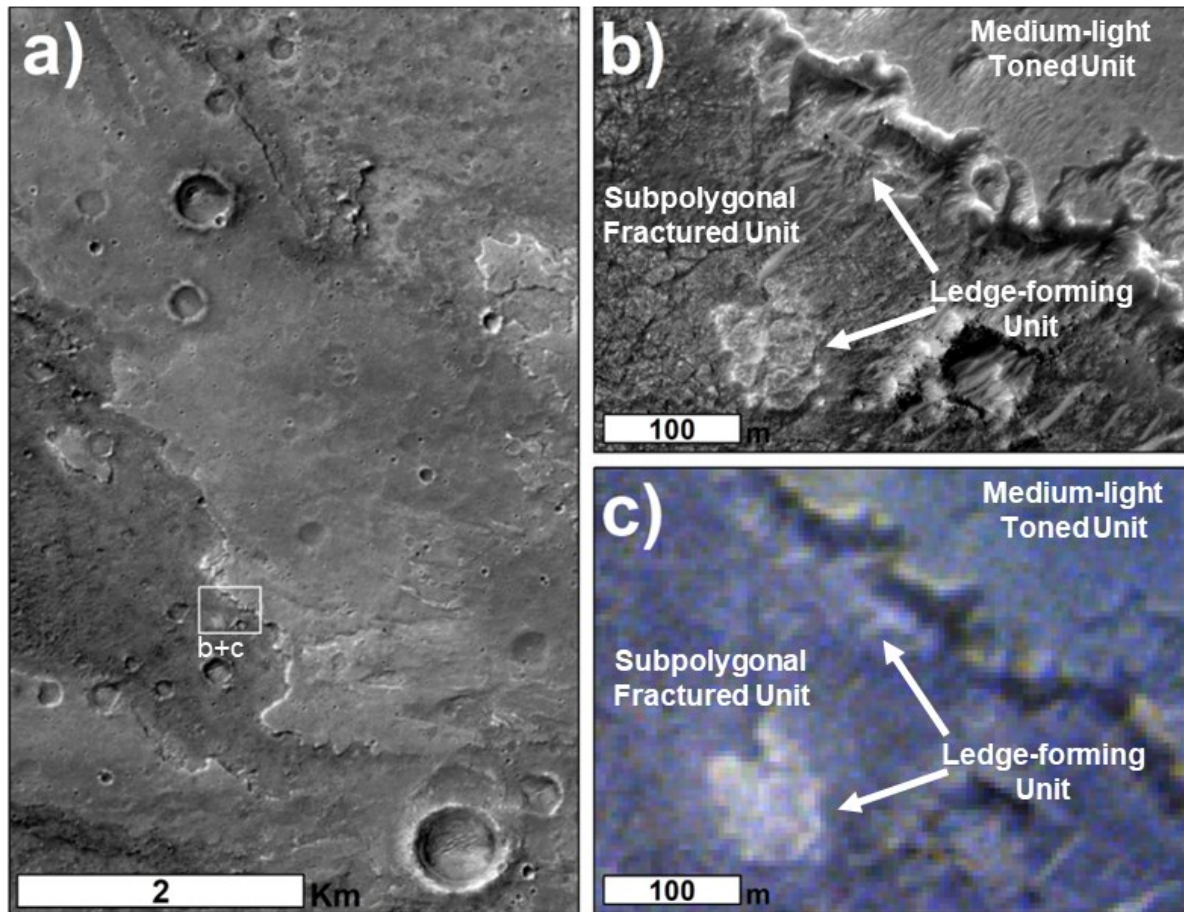


Figure 4.8: The Ledge Forming unit: (a) the context of b and c in the CTX mosaic; (b) the texture and tone of the Ledge-forming unit in HiRISE and its contact with the Subpolygonal Fractured unit and Medium-light Toned unit; (c) the texture and tone of the Ledge-Forming unit in CaSSIS.

4.4.2.3 Medium-light Toned Smooth Unit (Mlt-s)

Observations

The Medium-light Toned Smooth unit (Mlt-s) comprises five of the lobe-shaped bodies that form the sediment fan (see Section 4.5.2; Figure 4.17). The unit appears medium-light toned in both CTX and HiRISE images (Figure 4.9a-c), and medium-blue in the CaSSIS NPB images (Figure 4.9d). This unit is visually distinct from unit Mlt-r (Section 4.2.4), lacking the metre-scale knobbly surface roughness that characterises the latter and instead exhibiting a smoother texture with occasional fracturing (Figure 4.9b-c). In the central-eastern region of the fan's extent, unit Mlt-s typically forms steep scarps up to ~13 m high where it overlies the Basal units and underlying Le-f and LtF units. At the distal, basinward edge of the fan to the north-west, unit Mlt-s thins dramatically, tapering into <1 m thick exposures. The upper contact with the overlying unit Mlt-r is sharp, forming a well-defined cliff-bench morphology. The surface of unit Mlt-s is densely cratered, with many craters retaining fresh, well-defined rims, although circular ridges of possible ghost craters – buried craters – are also present.

At a m-scale scale, multiple examples of candidate stratification are observed (Figure 4.10). In two south-facing exposures of unit Mlt-s similarly proximal to the Coogoon Vallis outlet but approximately ~3 km apart, the unit is clearly layered. These approximately horizontal and planar strata can be traced continuously for 50-100 m (Figure 4.10c, d). The layering is internally comprised of dark- and light-toned appearing bands. The dark-toned bands (~1-2 m) lack any discernible sedimentary textures and are bracketed by the thinner light-toned bands that appear to be comprised of angular blocks. The light-toned bands occasionally end abruptly (Figure 4.10d) and also form resistant terraces (Figure 4.10c). Putative tabular cross-stratification is also identified within the flank of a ribbon-shaped ridge at the apex of one of the lobe-shaped sediment bodies (lobe-shaped body E; Section 4.5.2). This ~3 m thick flank exposes faintly discernible steeply inclined planar surfaces dipping to the west (Figure 4.10b).

The surface of unit Mlt-s occasionally exhibits sub-polygonal fracturing, dividing it into blocks 20-40 m across. Along the base of steeper scarps and within impact craters, rare clusters of light-toned boulders 2-5 m in diameter are also observed. Around the margins of unit Mlt-s, and draped across parts of the unit's surface, light-toned surficial ripple-forms are observed (Figures 4.9e, 4.10c).

Interpretations

While the fan's smooth appearance and low thermal inertia – the resistance of a material to temperature change (e.g. Melosh, 2011; Putzig and Mellon, 2007) – have led to previous interpretations that the sediment fan was unconsolidated (Ivanov et al., 2020), the retention of impact craters, expression of subpolygonal fracturing, and preservation of steep scarps indicate that unit Mlt-s is comprised of lithified rock (e.g. Edgett and Sarkar, 2021). Internal layering and the arrangement of the unit into lobe-shaped bodies support a sedimentary origin (e.g. Edgett and Sarkar, 2021; Malin and Edgett, 2000; Stack-Morgan, 2015). Previously interpreted clinopyroxene spectral signatures from CRISM spectral datasets across the fan surface (Platt et al., 2025; Turner and Bridges, 2017) would suggest a basaltic composition for the unit (e.g. Bedford et al., 2020; Rogers et al., 2018). However, surface alteration or a fine dust-covering have also been proposed to explain the spectral signatures observed in CRISM datasets at the same location (Ivanov et al., 2020). The distinctly low thermal inertia observed across the sediment fan (Ivanov et al., 2020; Quantin-Nataf et al., 2021) could be attributed to a sedimentary rock with high porosity, fracturing, and parting planes between laminae, as observed in Gale crater (e.g. Vasavada et al., 2017). Alternatively, the low thermal inertia could be explained by the breakdown of rock into grain-size particles in the uppermost centimetres of the fan's exposed surface, potentially supported by the light-toned, ripple forming deposits concentrated along the unit margins (e.g. Chojnacki et al., 2014). Occasional boulders are interpreted to be scree caused by fracturing.

The gently inclined planar dark- and light-toned appearing banding are interpreted to be metre-scale layering. Similar layering has been observed in sediment fans elsewhere on Mars (e.g. Morgan et al., 2022; Rice et al., 2013; Stack et al., 2020b) and is often attributed to subtle lithological differences between broad channel and flood plain deposits (Stack et al., 2024) common on fluvial fans (e.g. Gao

et al., 2020) and delta plains (Kleinhans, 2005). The discontinuity of one of the more resistant beds may reflect the lateral pinching-out of a broad channel body, though this remains unresolved at current resolution. The inclined geometry of the banding in the north-western portion of the fan could reflect cross-bedding or fracturing; distinguishing between the two is difficult from orbit. If cross-bedding, it may have formed by migrating fluvial bars, subaqueous dunes, or even as clinoforms forming in a shallow lake, if the sediment fan was indeed a delta. The presence of ghost craters across the unit suggests either partial burial or complete burial and partial exhumation by later depositional events (e.g. Arvidson, 1974; Roberts et al., 2021).

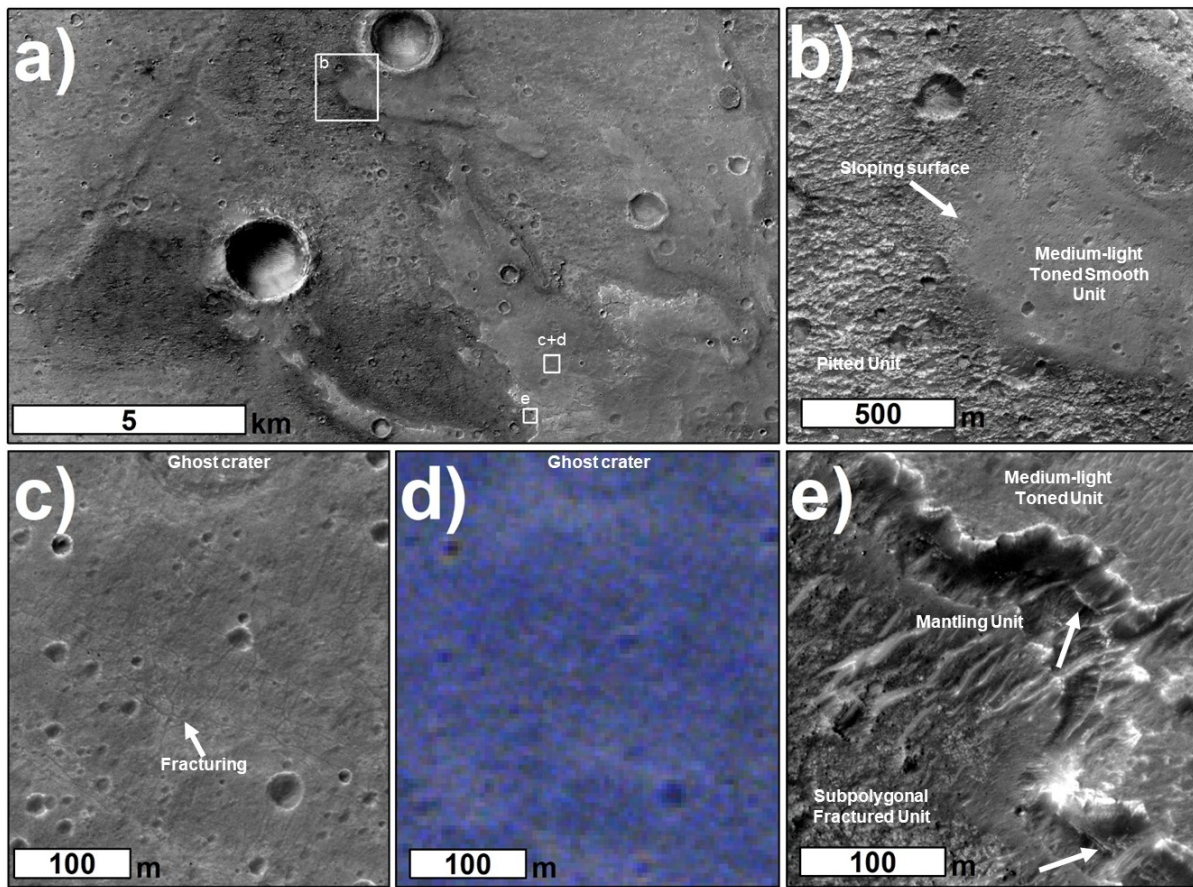


Figure 4.9: The Medium-light Toned Smooth unit: (a) the context of b, c, d, and e in the CTX mosaic; (b) the sloping erosional expression of the Medium-light Toned Smooth unit in the northern-most parts; (c) the texture and tone of the Medium-light Toned Smooth unit in HiRISE; (d) the texture and tone of the Medium-light Toned Smooth unit in CaSSIS; (e) the cliff-forming erosional expression of the Medium-light Toned Smooth unit. Arrows point out the occasional bench shape erosional expression.

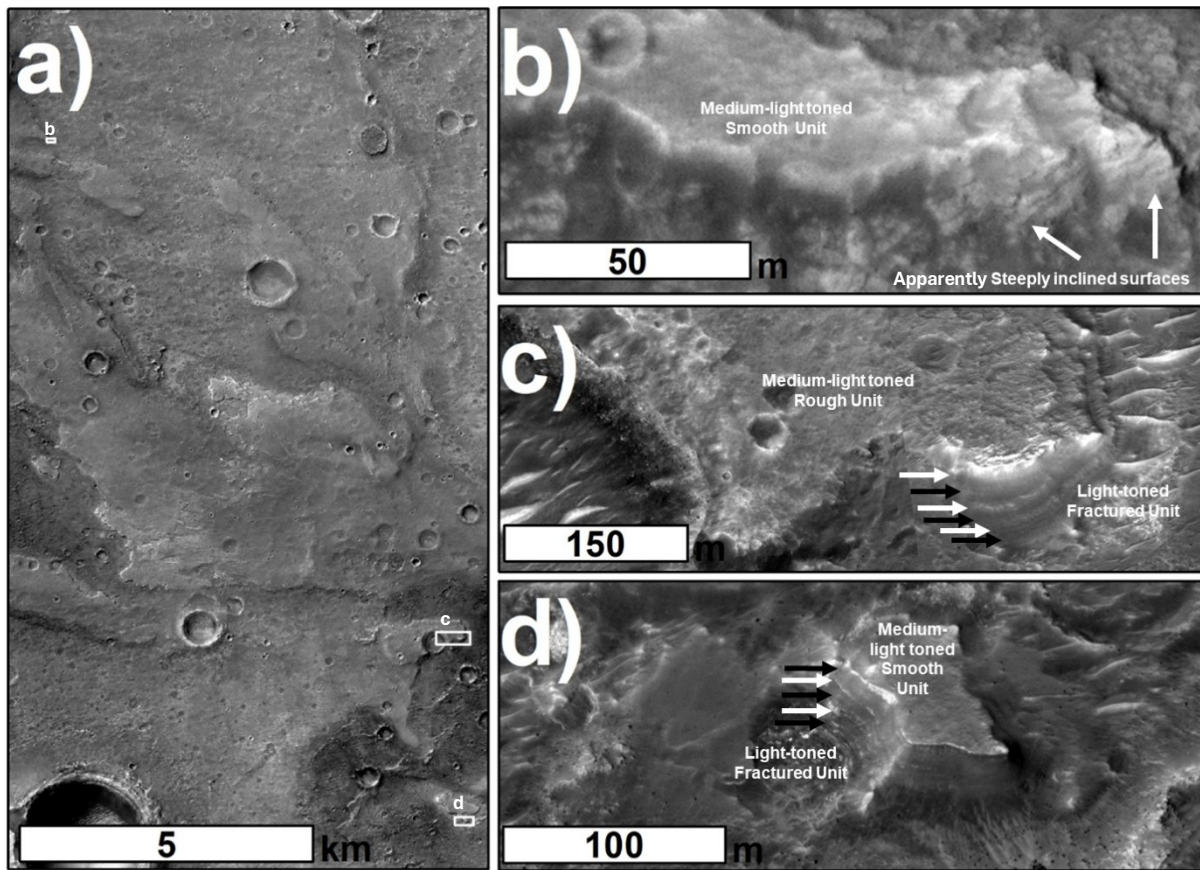


Figure 4.10: The Medium-light Toned Smooth unit: (a) the context of b, c, and d in the CTX mosaic; (b) apparently steeply dipping surfaces within the medium-light toned smooth unit within HiRISE; (c) gently inclined dark-toned and light-toned bands in HiRISE; (d) gently inclined dark-toned and light-toned bands in HiRISE.

4.4.2.4 Medium-light Toned Rough Unit (Mlt-r)

Observations

The Medium-light Toned Rough unit (MltR) overlies unit Mlt-S as a ~5 m thick triangular-shaped wedge that expands westward from Coogoon Vallis, terminating as an irregular scarp defined by linear ridges. Although the unit shares similar medium-light tones with MltS in CTX and HiRISE images (Figure 4.11a–c), unit MltR is distinguished by the lighter blue tones in CaSSIS NPB images (Figure 4.11d) and a knobby surface texture at the metre-scale (Figure 4.11c). The surface of MltR is densely cratered, with many impact craters retaining sharp rims. Ghost craters, inverted craters, and craters infilled with unit LtcF are also present (Figure 4.11c). In one example (Figure 4.14b and d), one of these craters is embedded within the Medium-light Toned Smooth unit and partially covered by the Medium-light Toned Rough unit. The northwestern and southern flanks of the unit, particularly near its western, basin-facing margin, are mantled by a light-toned ripple-forming, surficial deposit.

On a HiRISE scale, the most distinctive characteristic of MltR is its rough, knobby surface texture and linear ridges. The rough texture is characterised by scattered light-toned blocks or boulders ranging from 1–2 m in diameter that are not obviously associated with nearby scarps or impact craters. These

blocks or boulders are mostly concentrated in the eastern part of the unit, reaching ~30% surface coverage (Figure 4.11c), but their density progressively decreases westward to <1%. The finger-shaped linear ridges at the westward margin of the unit extend up to 500 m across the underlying unit. These features are straight, typically ~30 m wide, ~1 m thick, and occasionally record bifurcations. They are radially oriented with an average bearing of 292° (Fig. 13).

Fractures are rare within unit MltR, but one prominent example is a dark-toned, irregular, sharply defined groove 1–3 m wide, traceable for over 300 m across the unit's surface (Figure 4.12b). Additional subpolygonal fractures are observed in the northwestern part of the fan, confined to areas with smoother textures. These fractures partition the surface into ~50 m wide blocks, although they are less pronounced and more rare than the subpolygonal fracture networks observed in unit MltS. Finally, sets of parallel striations are etched into the unit's surface extending over 120 m and trending at a heading of 298° (Figure 4.12c).

Interpretation

The retention of impact craters, fracturing, and preservation of steep scarps indicate that unit MltR is at least partially comprised of cohesive rock (e.g. Edgett and Sarkar, 2021). However, due to the degraded appearance of scarps and the large volume of ripple-forming surficial deposits and talus which covers a higher surface area of this unit than the Medium-light Toned Smooth unit, the unit is interpreted to readily undergo break-down into grain-sized particles and boulders (e.g. Chojnacki et al., 2014) which may be the cause of the low thermal inertias (Ivanov et al., 2020; Quantin-Nataf et al., 2021) observed here. Similar to unit MltS, the arrangement of the unit into fan-shaped body support a sedimentary origin (e.g. Edgett and Sarkar, 2021; Malin and Edgett, 2000; Stack-Morgan, 2015) and clinopyroxene spectral signatures (Platt et al., 2025; Turner and Bridges, 2017) may suggest a basaltic composition for the unit (e.g. Bedford et al., 2020; Rogers et al., 2018).

From orbit, it is unclear whether the characteristic metre-size boulders that are distributed across the surface of unit MltR represent blocks that have broken down in-situ or represent an inherent sedimentary texture. Similar rough, knobby textured units have been identified on the Jezero fan from orbit (Stack et al., 2020b) and subsequently identified to be conglomerates from high-energy fluvial processes through rover observations (Caravaca et al., 2024; Mangold et al., 2024; Mangold et al., 2021; Stack et al., 2020b). The MltR unit may similarly represent a putative boulder-rich unit that could indicate high-energy fluvial processes compatible with both an upper delta plain and fluvial fan (e.g. Kleinhans, 2005; Weissmann et al., 2013). The gradual reduction in concentration of these protrusions from the proximal to distal part of this fan unit is then interpreted to be caused by a gradual change in the properties of this unit as this unit progressed basinward. Two hypotheses can be made: (1) there was a gradual increase in rock lithification progressing basinward – potentially through better cementation – leading to a reduction of in-situ breakdown; (2) or, if unit MltR did represent putative boulder-rich deposit, a gradual decrease in the size and distribution of boulders as the unit progressed basinward.

The linear ridges on the upper fan surface are interpreted to be inverted fluvial channels, formed where more erosion-resistant sediments were deposited within palaeochannels and subsequently exposed by deflation of the surrounding finer-grained material (e.g. Davis et al., 2019; Williams et al., 2009). Their width-to-thickness ratios (~30:1) align with the morphology of narrow fluvial sheets (Gibling, 2006).

The parallel striations may have formed via impact crater-generated winds (e.g. Schultz and Quintana, 2017) and the striations observed on the surface of the Medium-light Toned Rough unit are at an azimuth which points to Kilkhampton crater.

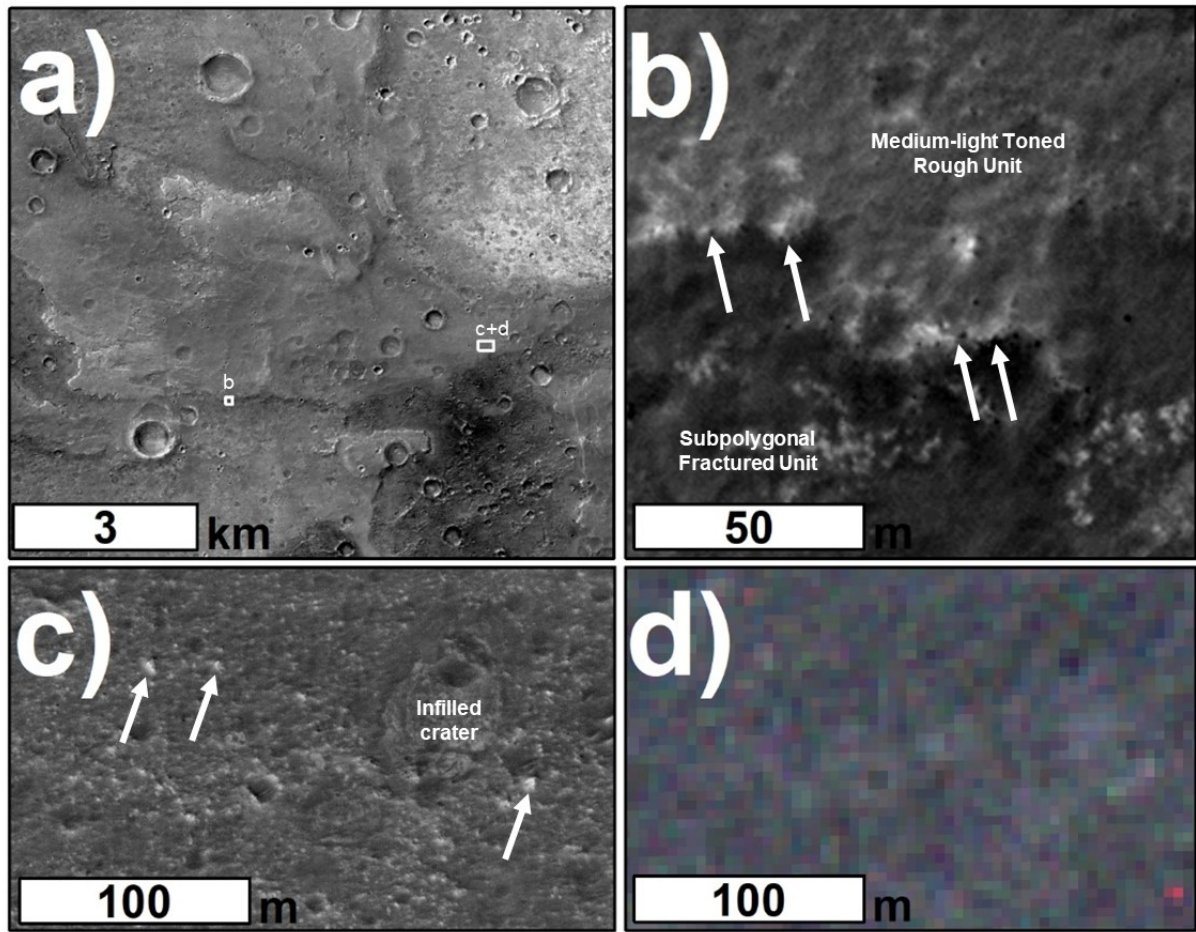


Figure 4.11: The Medium-light Toned Rough unit: (a) the context of b, c, and d in the CTX mosaic; (b) the erosional expression of the Medium-light Toned Rough unit and how the boulders/blocks are still exposed in cross-section (arrows point out prominent boulders within the unit scarp); (c) the texture and tone of the Medium-light Toned Rough unit in HiRISE with the Light-toned Crater Infill unit embedded within, arrows point to large boulders; (d) the texture and tone of the Medium-light Toned Rough unit in CaSSIS.

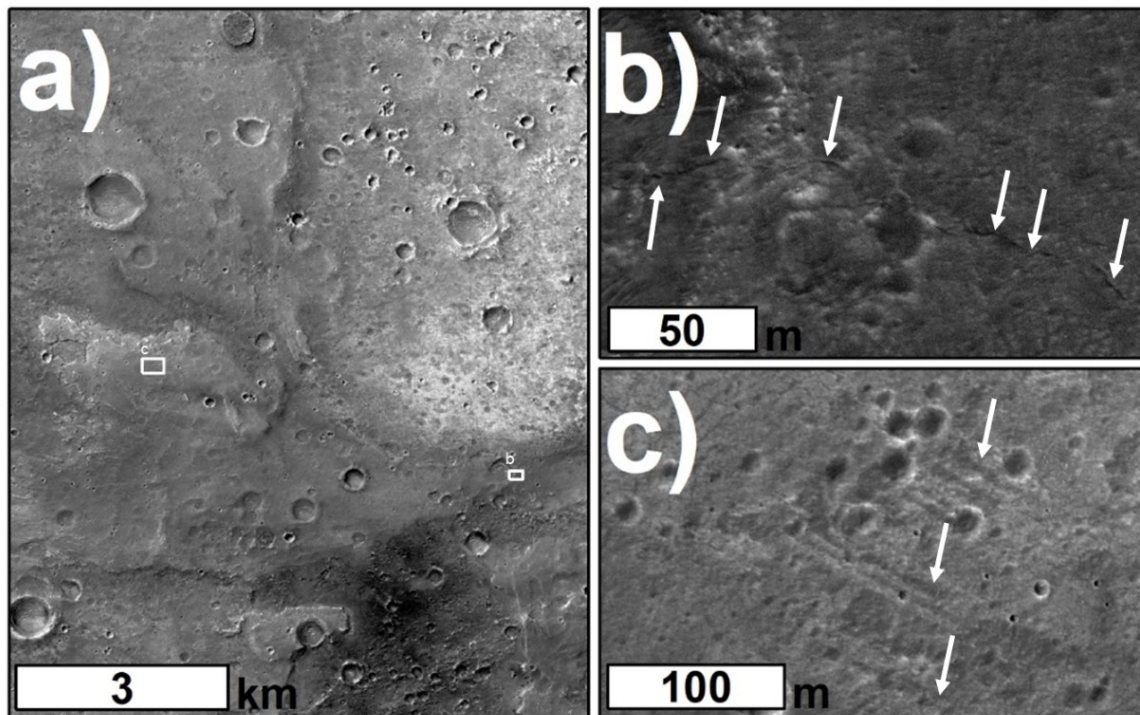


Figure 4.12: The Medium-light Toned Rough unit: (a) the context of b and c in the CTX mosaic; (b) a curvilinear, segmented fissure shown in HiRISE, arrows point out the segments of the fissure; (c) the striations in HiRISE.

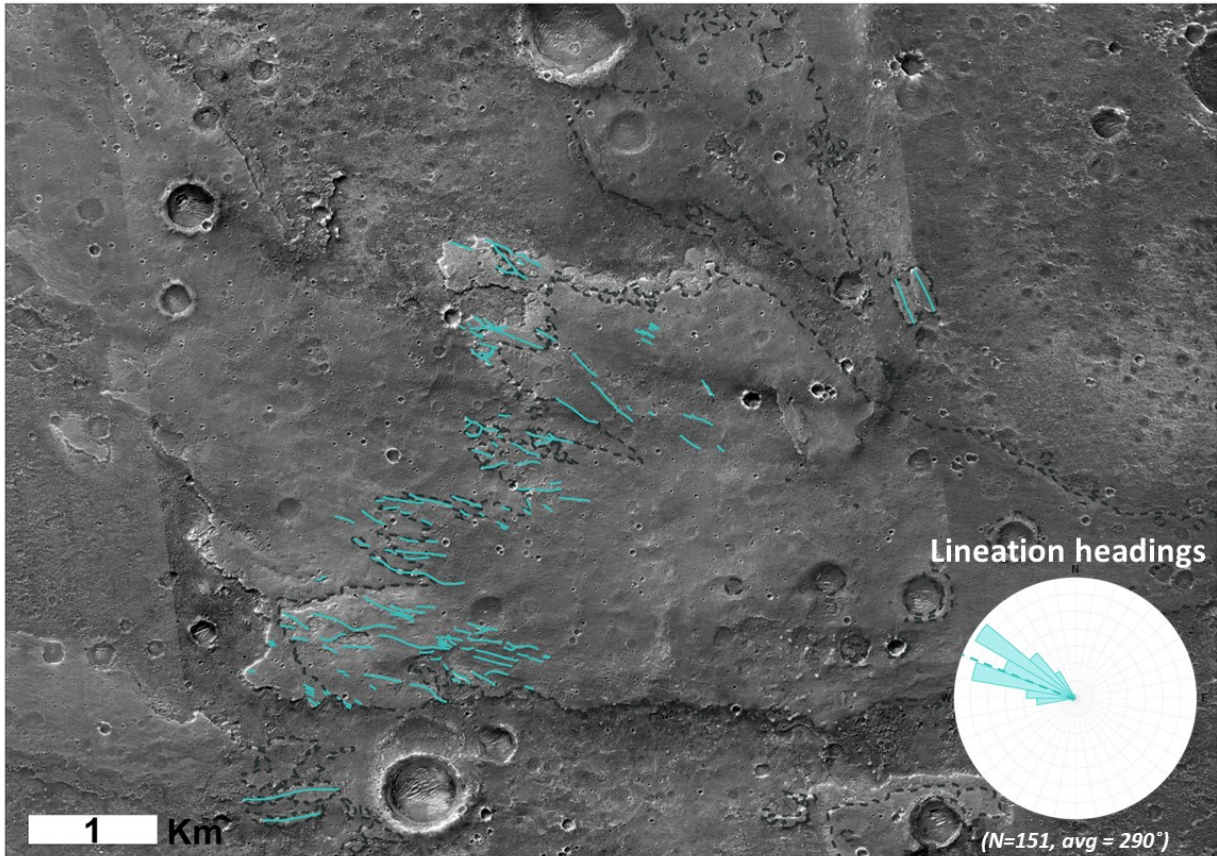


Figure 4.13: Linear ridges on the surface at the termination of the Medium-light Toned Rough unit. There are 151 lineations in total (turquoise) with their binned (every 10°) bearings shown as a rose diagram overlain by a dashed turquoise line representing the average heading.

4.4.3 Other Units

4.4.3.1 Light-toned Crater Infill Unit

Observations

The Light-toned Crater Infill unit is restricted to small, circular depressions - 20 to 250 m in diameter – and are characterised by high albedos in CTX (Figure 4.14a) and HiRISE (Figure 4.14b, d) images, and a distinct white-blue tone in CaSSIS images (Figure 4.14c, e). The surface texture of this unit is distinct for its thin, circular, light- and dark-toned bands (1 – 5 m wide) that are parallel to the depression rim. These features are primarily found embedded within the Medium-light Toned Rough unit with up to 25 individual occurrences. Other examples are also found within depressions in the Lower Fractured unit. In two examples (Figure 4.14b and d), these craters are embedded within the Medium-light Toned Smooth unit and partially covered by the Medium-light Toned Rough unit. This unit is crater retaining and retains craters as small as ~3m and as large as ~20m in diameter.

Interpretation:

These light-toned in-filled craters are widespread in Oxia Planum (Fawdon et al., 2024) and are

likely caused by a regional process. Processes such as ash deposits infilling a crater and subsequent sagging may cause a concentric pattern (Arp et al., 2021). Or, alternatively, this pattern of concentric rings have been found in satellite remote sensing on terrestrial lakes caused by slow evaporation processes (Abbas et al., 2018).

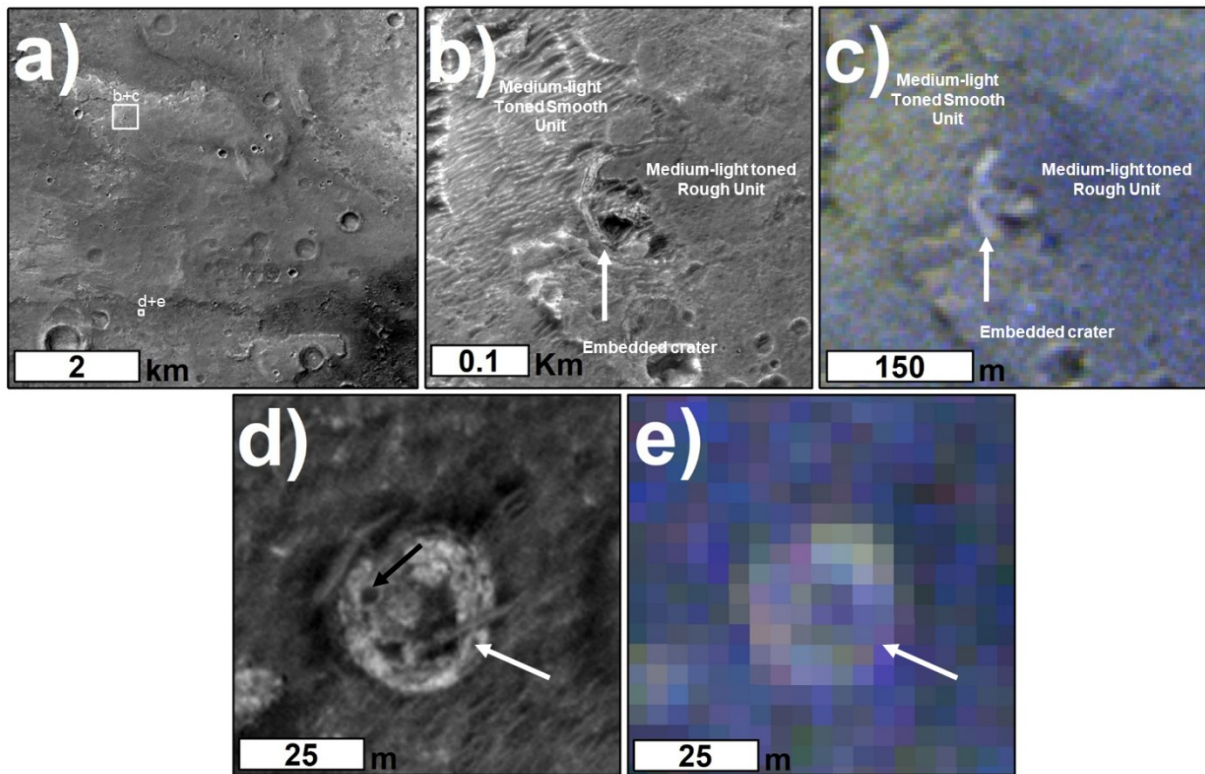


Figure 4.14: The Light-toned Crater Infill unit: (a) the context of b, c, d, and e in the CTX mosaic; (b) the embedded crater between the medium light-toned smooth and rough units in HiRISE; (c) the embedded crater between the medium light-toned smooth and rough units in CaSSIS; (d) example of uncovered light-toned crater infill in HiRISE (black arrow points to crater within infill unit); (e) example of uncovered light-toned crater infill in CaSSIS.

4.5 Morphology of the Sediment Fan

4.5.1 Channel-Basin Transition

The channel-basin transition is difficult to identify; there is a gradual change in elevation from the areas of higher relief to the east, to where the channel is incised to the basin containing the sediment fan. The -2950 m contour line generated from the CTX DTM appears to best demarcate the point where the fluvial system becomes unconfined. This transition has likely been eroded back and changed even after fan emplacement, so the channel-basin location is used as an arbitrary placeholder for further documentation of fan geomorphology and interpretations.

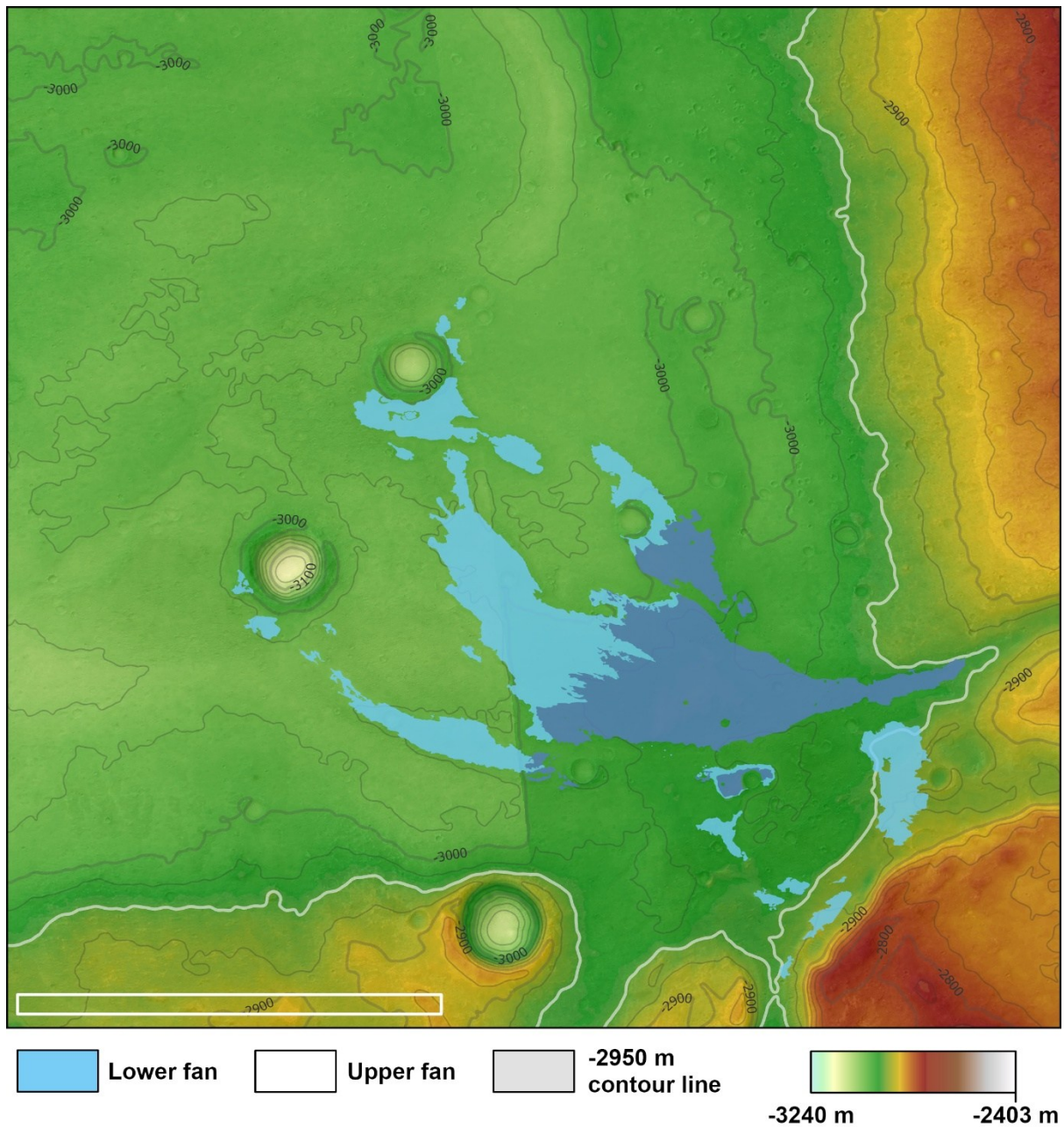


Figure 4.15: The arbitrary placeholder -2950 contour line used to determine the channel-basin transition in white. The lower fan (dark blue) and upper fan (light blue) are also shown against the blended CTX MOLA DTM against the orthorectified CTX mosaic.

4.5.2 Ribbon and Lobe-shaped Sedimentary Bodies

Observations

The sediment fan at Oxia Planum covers a total surface area of 40.5 km² and consists of seven finger-shaped, lobe-shaped and wedge-shaped sedimentary bodies. These sedimentary bodies occur between elevations of -2965 m and -3039 m with a maximum recorded thickness of ~5 m relative to the immediately adjacent underlying stratigraphy (this measurement deliberately did not take into account the north-south orientated ridge cross-cutting the fan stratigraphy which was unlikely a depositional

feature; section 4.5.5). These bodies were identified in HiRISE orthomosaics and the HiRISE and CTX blended DTMs and given letters from A to G (Table 4.2; Figure 4.17). As the sediment fan is highly eroded and cratered, only parts of the sediment lobes have been preserved and the relation between nearby sediment lobes is sometimes unclear (for example between sediment bodies D and F). Therefore, there are some uncertainties in either the former full extent of the sediment body, the shape of the sediment body, and the relation to other sediment bodies. In this section, the morphology of these bodies and their relationship to other bodies will be described and interpreted.

The triangular-shaped deposit, sediment body G, is better preserved than the nearby sediment lobes for its retained ridges and how it expands in width westward from Coogoon Vallis terminating over sediment body B as an irregular scarp with linear ridges. Isolated outcrops have been identified as part of this sedimentary body due to surface texture (Section 4.4.2.4) which overly sediment bodies E and F.

Sediment bodies D, E, and F are disconnected from the main sediment fan body and vary in morphology. Sediment body F begins at the termination of sediment body G as an elongated lobate shape. Sediment body E occurs as an elongated, curved body reaching widths of ~990 m at its western end before narrowing to widths of 380 m. Sediment body E terminates in a series of isolated outcrops reaching a total width of 897 m. Sediment body D is preserved in multiple remnants with a short ribbon-like ridge (~640x1890 m), which then expands north-westward into a triangular deposit reaching widths of ~3,820m.

Sediment body C is identified by a scarp visible in both the HiRISE and CaSSIS datasets which truncates to the west over sediment body A. Sediment body C begins as a ribbon-shaped body ~660 m in width at its thickest and ~70 m at its narrowest, extending north-westward for ~2490 m. At the end of the ribbon-shaped body, the lobe terminates as a series of isolated outcrops forming a westward expanding triangular shape. This triangular shape abruptly ends at the southern side of sediment body D. While there is a m-scale gap between the two sediment bodies, the truncated margin of sediment body C is a geometric fit for the southern margin of sediment lobe D.

Sediment body B is the largest preserved sediment body, though appears to be heavily degraded and does not appear to preserve its original form. This lobe is overlain by sediment body G. The southwestern margin appears as sharp digitate extensions which partially overly a crater embedded within sediment body A. The boundary between sediment body A and sediment body B is obscured towards the north-eastern side of this unit in HiRISE images, however, there is a distinct change in elevation observed in the HiRISE blended DTM at an elevation of around -3001m.

The north-western side of lobe-shaped body A truncates over the basin as an irregular, digitate scarp. These finger-like-shaped extensions extend for up to ~510m at consistent widths of between ~110m and ~140m.

Sediment Body	Area (km ²)	Headings (°)	Elevations (m)
A	4.07	293-327	-3038 to -2996
B	5.29	300-294	-3034 to -2970
C	1.31	~326	-3030 and -2998
D	4.02	~296	-3026 and -2993
E	3.19	~297	-3039 and -2983
F	3.27	~324	-3016 and -2974
G	17.6	245-330	-2996 and -2946

Table 4.2: Quantitative data for sediment bodies A through G.

Interpretations

The ribbon-shaped sediment bodies are interpreted to be the remnants of exhumed and topographically inverted feeder channels. At their terminus, the sediment fan bodies are interpreted as the remnants of inverted fan lobes, where a network of distributaries formed around an avulsion node (e.g. Bhattacharya, 2006; Moscariello, 2018; Weissmann et al., 2013).

Due to the cross-cutting relationships, a depositional order can be proposed for each sediment fan lobe. Previous attempts to determine the depositional order for other sediment fan lobes on Mars have used elevation as an additional tool to determine the order of emplacement (Stack et al., 2020a). As there is evidence for a high degree of erosion in the sediment fan at Oxia Planum (Section 4.4.2), maximum and minimum elevations of the sediment fan lobes will not solely be used to determine the order of emplacement (which is shown in Figure 4.17).

As no units truncate or cross-cut sediment body G and sediment body G superposes sediment bodies D, E, and F, sediment body G is interpreted to be the youngest lobe-shaped sediment body in the mapping area. As sediment bodies D, E, and F are isolated from the sediment fan, their cross-cutting relationships to each other are unidentifiable. However, the geometric fit of the northwest margin of sediment body C to the southwest margin of sediment body E, likely indicates sediment body E superposed sediment body C. Both sediment bodies B and C have margins which superposes sediment body A. Therefore, sediment body A is interpreted as the oldest fan deposit within the order of emplacement.

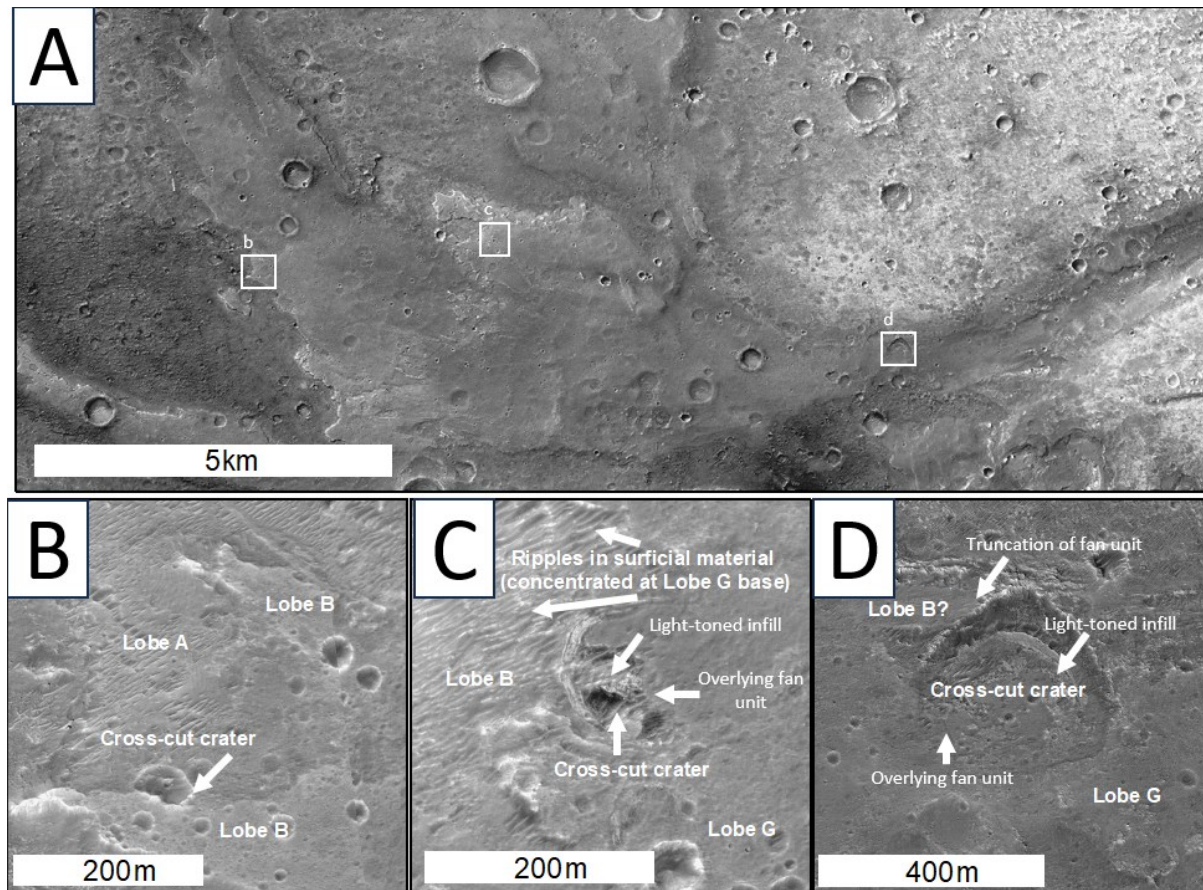


Figure 4.16: A) the location of the three cross-cut craters from sediment lobes in CTX. B) The location of a crater embedded within Lobe A, partially overlain by Lobe B. C) An infilled and lithified crater embedded within Lobe B which is partially overlain by Lobe G. D) A crater truncating an underlying fan unit which is infilled with a light-toned material, lithified, and partially overlain by Lobe G.

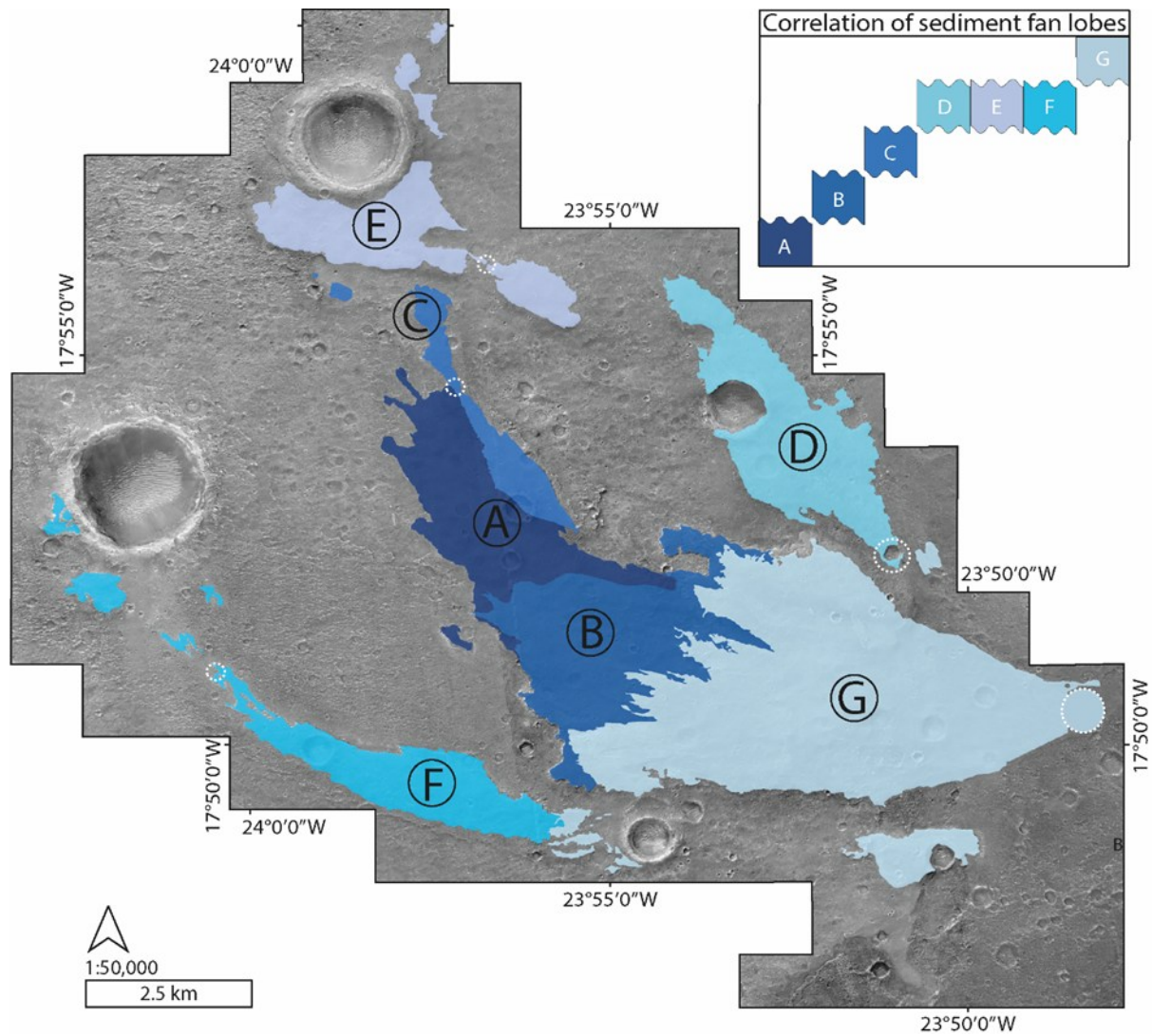


Figure 4.17: Each of the sedimentary bodies and their corresponding alphabetical letter against the orthorectified HiRISE mosaic with correlation of map units. Interpreted expansional apices are delineated by dashed white circles.

4.5.3 The distribution of proximal apices and their distance to the channel-basin transition

Observations

The distance of the apex at which the ribbon-shaped sedimentary bodies expand into a lobe-shaped sedimentary body is described here in respect to the distributions and distance from the arbitrary channel-basin transition. As the channel-basin transition likely changed over the construction of the sediment fan and has faced subsequent erosion making the transition difficult to identify, the distances for the expansional apices to the channel-basin transition are not absolute values, but instead a reference point to compare the distances of the sediment body apices relevant to each other.

The youngest sediment body, sediment body G, commences with a ribbon-shaped body incised into the highlands of the mapping area which transitions into a triangular-shaped body with an apex at the

channel-basin transition (Figure 4.15) therefore having an apex ~ 0 m from the channel-basin transition. Sediment bodies E and F both expand from ribbon-shaped sedimentary bodies into remnants of lobe-shaped sedimentary bodies at around $\sim 11,000$ m and $\sim 13,000$ m north-west from the channel-basin transition respectively. Sediment body C expands into a lobe-shaped apex at $\sim 10,500$ m from the channel-basin transition. Finally, as sediment bodies A, B, and D are covered by sediment body G, the expansional apices have been obscured but can be reasonably approximated to have occurred $< 9,000$ m, $< 6,500$ m, and $< 6,200$ m from the apex respectively.

Interpretations

The earliest expansional apices of the sediment fan occur close to the channel-basin transition. The expansional apices then jump basinward over sediment body A by at least $\sim 5,400$ m and remain around the modern-day elevation of ~ 3025 m for multiple periods of lobe deposition. Finally, the lattermost lobe of the sediment fan jumps back channel-ward and returns to the channel-basin transition.

4.2.4 Sediment Fan Slope

Although it is difficult to constrain a mean slope for the lower units as these are heavily degraded to the point where the lobes are preserved only as remnants, lobe G is relatively intact with most of its lineations preserved and is therefore a candidate for measuring the slope of the fan. The slope of this lobe was measured (Figure 4.18) and was found to show a mean slope of 0.1° . There is a large vertical offset in the data at around $2000 - 3000$ m from the expansional apex of the lobe of approximately ~ 10 m visible not only in the mean topographic profile but in all the topographic profiles.

Interpretation

The topographic profile of the fan shows only a slight downward slope and remains comparatively flat, a form more consistent with fans in fluvial or deltaic environments and distinct from the steeper gradients typical of alluvial fans (Blair and McPherson, 1994, 2009). The notable amount of displacement that occurs midway across lobe G is interpreted as being connected to the kilometre-scale ridge that extends through the area (Section 4.5.5).

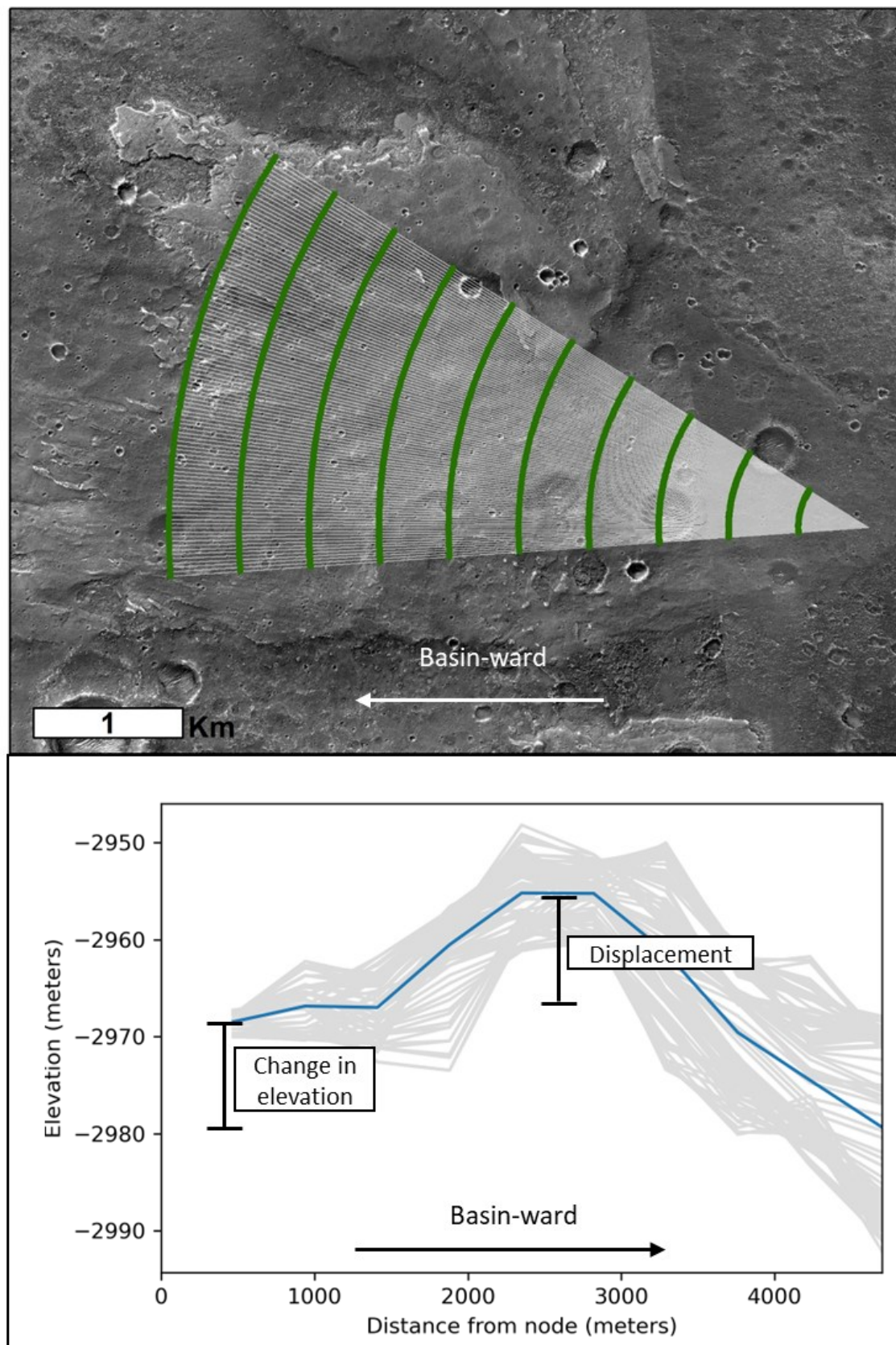


Figure 4.18: a) The grid that the average slope is taken from. The 100 equally space lines between the specified headings are shown in white and the points at which the elevations are taken are shown in green. b) The elevation profile of the upper sediment fan. The mean elevation profile is blue whilst all elevation profiles taken are light grey.

4.5.5 The North-South Ridge

Previous CTX-scale mapping attempts identified an additional elongated lobe-shaped sediment body at a heading of $\sim 8^\circ$ extending for $\sim 1630\text{m}$ (Quantin-Nataf et al., 2021). This ridge is significant in the blended CTX DTM (Figure 4.19a) reaching elevations of around -2976 m compared to the elevation of around -3000 m of the basin floor immediately to the East of the ridge. However, when investigated using the HiRISE and CaSSIS datasets, the ridge shared a texture and tone similar to the Orthogonally Fractured unit (Section 4.4.1) and did not share any characteristics with the Medium-light Toned Smooth and Rough units (Section 4.4.2). When investigated outside of the mapping area, this ridge, although reduced in elevation to -2993 m compared to the surrounding -3000 m basin floor, curved around to a heading of $\sim 349^\circ$ in line with a prominent ridge extending at the same heading for $\sim 14,210\text{ m}$ at elevations of around -2960 m compared to the -3000 m surrounding basin floor.

A similar ridge (Figure 4.19b) was found within the Medium-light Toned Rough unit at a heading of ~ 6 degrees extending $\sim 980\text{m}$. When the average slope of the Medium-light Toned Rough unit was collated, this ridge appeared prominently showing that it was significant enough to impact the entire Medium-light Toned Rough unit displacing the unit by $\sim 10\text{ m}$ (Figure 4.18). This larger scale extended ridge is parallel to a similar ridge which is $\sim 43.5\text{ km}$ in length with an average heading of ~ 354 degrees around 17 km away from the first (Figure 4.19a).

Interpretation:

This ridge is interpreted to be an extension of a large-scale fault. Previously, compressional fault regimes have been mapped in Oxia Planum with a similar average heading between NNW and NNE (Woodley et al., 2021). As this ridge displaces the sediment fan, it is interpreted to occur after the sediment fan was emplaced. The additional ridge to the East of the mapping area is cross-cut by the feeder channel (Figure 4.19c) and I therefore interpret this episode of tectonic activity to occur before the emplacement of the upper fan unit. This finding supports previous interpretations from Woodley (2021) that tectonic activity occurred in multiple stages in Western Arabia Terra.

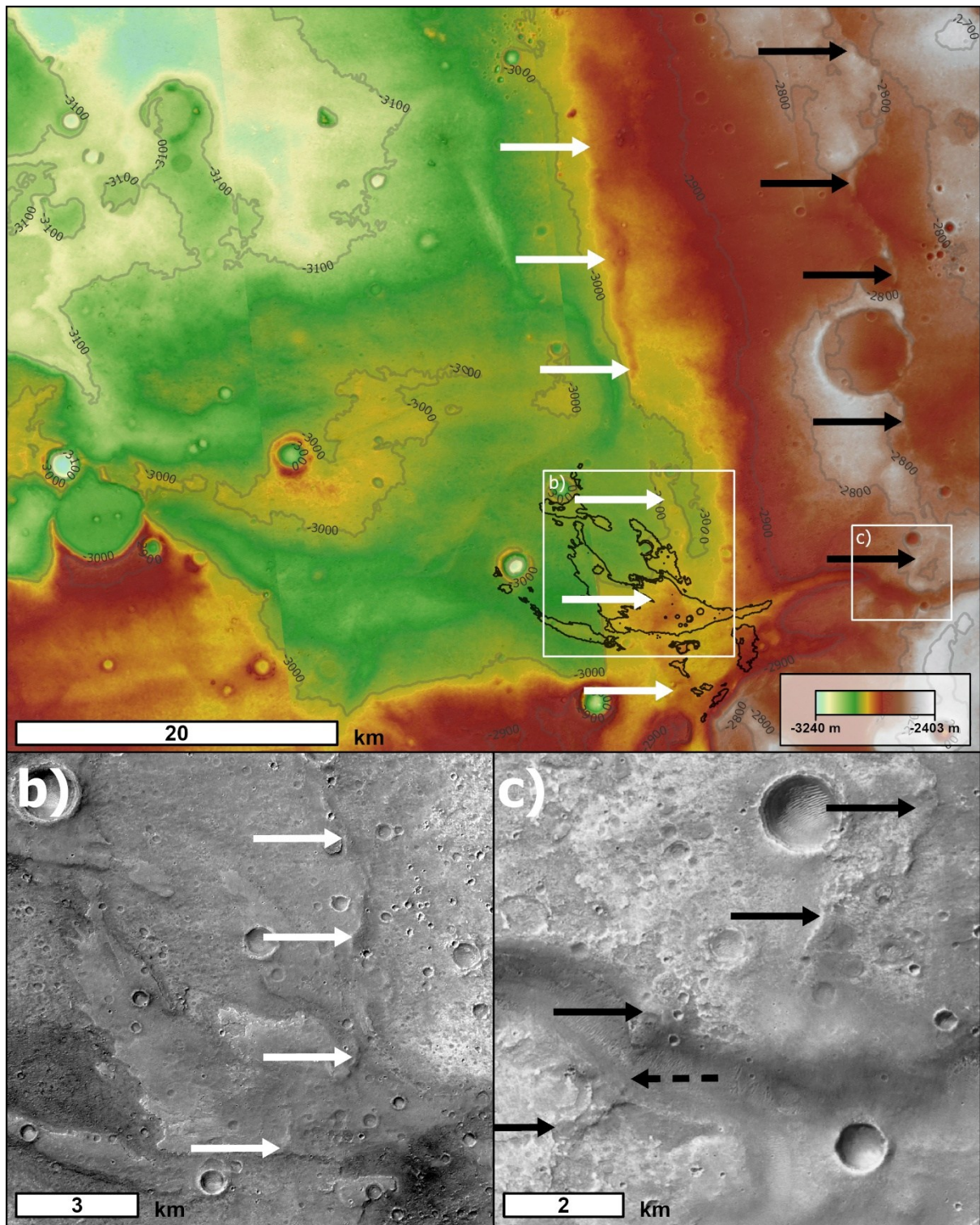


Figure 4.19: North-south orientated ridges: (a) the ridges in the CTX blended MOLA DTM against the Orthorectified CTX mosaic with the ridges pointed out in black and white arrows and the context of b and c; (b) the eastern ridge which cuts through the sediment fan (white arrows) against the orthorectified CTX mosaic (c) the western north-south ridge (black arrows) which is cross-cut by a source channel (shown with the dashed black line) against the orthorectified CTX mosaic.

4.6 Discussion

4.6.1 History and timing of Emplacement

Outcropping at the lowest elevations, units RI and LF are interpreted as the basement rock of the study area and the lowermost units within the stratigraphic column (Figure 4.2). These units correspond to the previously mapped layered and fractured Phyllosilicate-bearing unit (e.g. Ivanov et al., 2020; Mandon et al., 2021; Quantin-Nataf et al., 2021). Ivanov et al. (2020) reported an age of 3.53 Ga through crater counting and thus found this unit originated from the Noachian and Hesperian boundary similar to Tanaka et al. (2014). The phyllosilicates in this unit are attributed to either deposition within a subaqueous environment (authigenic or detrital in origin) or through subsequent aqueous alteration of the surface or near-surface (Mandon et al., 2021; Quantin-Nataf et al., 2021). Although desiccation and hydrofracturing could cause the planform polygonal patterns observed in the phyllosilicate-bearing unit, other research have attributed fracturing to have occurred at a later date through processes related to burial (e.g. Bowen et al., 2022; Quantin-Nataf et al., 2021). While this unit is presently exposed at the surface, McNeil et al. (2021) have identified that ~40 m (locally reaching up to ~130 m) of vertical stratigraphy have been removed from Oxia Planum and thus fracturing related to burial is a possibility.

The raised linear features of unit RI are similar to a ‘boxwork’ texture observed elsewhere on Mars via satellite images (e.g. Anderson and Bell III, 2010). Subsequent rover surface studies of similar features have found these structures can occur through cementation of fractures from groundwater (e.g. Kronyak et al., 2019) or through localized aqueous alteration through soft-sediment likely as ice-wedges (Banham et al., 2024b). Subsequent differential erosion would have removed the surrounding material leaving the more erosion-resistant fracture-fill. As these raised linear features can be traced laterally across to the deep fractures of the Lower Fractured unit (Section 4.4.1.1; Figure 4.4d), the two processes are interpreted to be related, and a post-depositional and burial hypothesis is favoured. Thus, these raised linear features are attributed to form after phyllosilicate formation and fracturing of the Lower Fractured unit but before current-day erosion.

Units P and Sf occur stratigraphically above unit LF and are interpreted to be emplaced coevally. These units correspond to the Dark Resistant unit where crater counting suggests an Amazonian age (Quantin-Nataf et al., 2021). However, the pitted and degraded nature of these units suggest that many craters may be difficult to distinguish in crater counting. The rough texture of the unit P is interpreted to be an eroded form of the unit Sf due to the gradational contact, increase in pits, and frayed edges (Section 4.4.1.4; Figure 4.5 and Figure 4.6). Unit Sf is found closer to the sediment fan and is therefore interpreted to have been protected by the sediment fan either by the raised elevation or more likely because the sediment fan has been eroded back. The steep bench-shaped contact between this unit and the overlying sediment fan puts this unit emplacement before the sediment fan was deposited in accordance with principles of photographic recognition of sedimentary rocks developed for Mars

(Edgett and Sarkar, 2021). This interpretation is at odds with previous work, which puts the pitted unit forming after the sediment fan was deposited based on apparent draping relationships and crater counting which placed this unit as Amazonian in age (Quantin-Nataf et al., 2021). However, the interpretation here is consistent with other studies, which finds that the sediment fan post-dates the pitted unit (e.g. Gary-Bicas and Rogers, 2021).

As the sediment fan in Oxia Planum can broadly be divided not just by the lobe-shaped features but also by the surface texture, this study supports previous work showing that this fan had multiple episodes of sediment deposition (Quantin-Nataf et al., 2021). Three craters ranging from 75-300 m have been found to be embedded within layers of the sediment fan with a smaller crater between two of the lobe-shaped bodies (sediment bodies A and B) (section 4.5.2; Figure 4.16), providing evidence for a small hiatus in deposition between the emplacement of these bodies. The large, infilled crater between the Medium-light Toned Smooth unit and Medium-light Toned Rough unit (Figure 4.14) likely indicates a longer hiatus in deposition as this crater was not only impacted into the Medium-light toned Smooth unit before the Medium-light Toned Rough unit was deposited, but it was also infilled and lithified before sediment fan reactivation.

The material which infilled the craters between the Medium-light Toned Rough and Smooth units was likely a regional process found throughout Oxia Planum, while the material is unknown previous studies of concentric ring infillings have been caused by ash deposits and evaporites on Earth (e.g. Abbas et al., 2018; Arp et al., 2021). Due to the size of these craters between the Medium-light Toned Rough and Smooth units, Quantin-Nataf et al. (2021) suggested that the temporal gap between the two units could be up to millions of years.

The Medium-light Toned Rough unit is the last ‘bedrock unit’ deposited in this area and likely indicates a change in environment due to the change in texture. Striations on this unit similar to striations caused by impact generated winds (e.g. Schultz and Quintana, 2017) were found at an azimuth which pointed to the nearby Kilkhampston crater. If these were striations from impact generated winds, this could be an additional line of evidence which indicated this unit pre-dates Kilkhampston crater. This supports findings from Gary-Bicas and Rogers (2021) and Fawdon et al. (2022) where the source channel for the sediment fan is covered by a layer of ejecta from the nearby Kilkhampston crater. Crater counting Kilkhampston crater’s ejecta blanket has found that Kilkhampston crater is likely around 3.67 Ga (Gary-Bicas and Rogers, 2021) putting an upper limit on the formation of the sediment fan.

There has also been tectonic activity in the region (Woodley et al., 2023; Woodley et al., 2024). This tectonic activity may have occurred both prior and after the deposition of the sediment fan as the source channel cross-cuts one of the tectonic ridges and one of the tectonic ridges cross-cuts the upper fan unit.

4.6.2 Towards a Working Hypothesis on the Fan Emplacement Model

The seven sediment bodies in Oxia Planum indicate evidence for at least seven different episodes

of sediment deposition with two significant hiatuses, inferred from the presence of embedded craters between three sediment fan layers. As the Medium-light Toned unit and the Medium-light Toned Rough unit have varying surficial textures, degrees of lithification, morphological character and a temporal gap, their morphologies and what can be inferred about the potential palaeoenvironments will be discussed separately. The distinction in shape and morphology of sediment body G, comprising unit MltR, and the hiatus represented by the infilled crater, may indicate a different depositional environment, potentially a fluvial fan or a delta top, overlying pre-existing fan deposits.

4.6.2.1 Lower Fan

Unlike many well-preserved examples of sediment fans on Mars (Fawdon et al., 2018; Goudge et al., 2017; Rice et al., 2013; Stack et al., 2020a), the lower fan is highly degraded and does not contain any lineations which can be interpreted as relict channels. Despite this, there are still preserved sedimentary features which can aid in the interpretation of the palaeoenvironment such as tenuous, metre-scale exposed layering within outcrop scarps (Figure 4.10). The dip of the inclinations of these strata do appear to change basin-ward, from horizontal to shallowly dipping interbedded dark-toned and light-toned strata at locations in the scarps closer to the source channel to more steeply inclined interbedded dark-toned and light-toned appearing strata basin-ward just at an expansional node (Figure 4.10). Remote sensing techniques have previously used the stratigraphic architecture of layers within fan deposits to make inferences about the depositional environment. In Jezero, previous orbital research interpreted steeply dipping layering that appeared to downlap basinward into shallowly inclined layers as clinoforms in a delta system (Goudge et al., 2017). These candidate clinoforms were subsequently corroborated by later in-situ imaging by the Perseverance rover (Mangold et al., 2021). Although the Oxia Planum sediment fan does appear to have shallowly dipping layering, which could be interpreted to be topsets, and inclined layering which could be interpreted to be foresets, (Figure 4.10) the layers within the Jezero fan have greater exposure and are larger and thus could be fitted with planes and mapped. For the exposed layering seen within the Oxia Planum sediment fan, the resolution of the apparent strata, being at the sub-metre-scale, the small amount of exposure, and the noise within the HiRISE DTM makes plane fitting difficult and thus tenuous to call the change in the layering as clinoforms.

The lower fan, distinguished by the Medium-light Toned Smooth unit, encompasses lobe-shaped sediment bodies A to F (section 4.5.2). The expansional apices of these lobes, the point where the channel ‘expands’ from the inverted channel into the lobe (Figure 4.17), are interpreted to be remnants of avulsion nodes. Moving up the stratigraphy from sediment body A to F, the nodes jump basinward over 5,400 m forward and shift laterally. These apparent shifts do not appear to be caused by changes in topography with the basin being relatively flat with no substantial changes in slope. There is also no evidence of distinctive channel-cutting creating new valleys and valley mouths, a method that has been found to cause the rare prograding of fluvial fan avulsion nodes in the past (Davidson et al., 2013;

Weissmann et al., 2013). Therefore, there appears to be migration (and migration laterally) of avulsion nodes moving basinward independent of topographic controls. This basinward migration, therefore, may be caused instead by the changing backwater-mediated zone found in fan deltas abutting into a shallow body of water (Van Dijk et al., 2012). This phenomenon has been observed remotely on Earth in the Volga and Omo delta to find that downstepping nodes is often caused by retreating shoreline levels and characterised through the elongated nature of the lobes (Reynolds, 2022). This has also been observed on Mars as Hypanis Valles was also found to have prograding avulsion nodes with no observed channel-cutting relationships which were also likely caused by shoreline retreat (Fawdon et al., 2018). Like with Hypanis Valles, due to the elongated-appearance of these lobe-shaped sedimentary bodies and the apparent basinward migration of nodes, the lower fan could be consistent with a deltaic model.

Alternatively, the lower fan also resembles fans formed under sustained, low-sediment discharge over extended periods of time (Hartley et al., 2010). Comparable fans on Earth can take thousands to tens of thousands of years to form (e.g. Donselaar et al., 2022), suggesting the lower fan may have been a similarly long-lived system, consistent with inferred ages for nearby fluvial systems in Arabia Terra that also formed in the mid-Noachian to early Hesperian (Davis et al., 2019).

4.6.2.2 Upper Fan

From orbit, so far only a small number of Martian fan systems, including Jezero, show evidence of recording regional climatic changes (see Section 5.2.2.2). These systems have deposits with distinct photogeologic textures that are superimposed over layered, comparatively smoother sediment-fan systems (Figure 4.3). These superimposed deposits are typically characterised by a rough appearance with metre-scale blocks or boulders and lack internal structures or layering. Reported examples include Jezero (Goudge et al., 2018), Holden (Grant et al., 2008), among other locations (Grindrod et al., 2018; Pondrelli et al., 2008; Williams et al., 2011).

The geomorphology of the upper fan unit can help unveil the depositional palaeoenvironment of the fan at the time of emplacement. Unlike the lower fan unit, the upper fan unit is too degraded to identify internal stratification so internal layering cannot be used as an identifier for processes which formed the upper fan. However, the basinward portion of the upper fan is characterised by raised lineations with azimuths spread in a radial, albeit bimodal, fashion (Figure 4.13). These lineations are interpreted to be inverted channels which are found elsewhere on Mars and in Oxia Planum itself where channels were previously infilled with a more erosion-resistant material than the surrounding strata (Davis et al., 2023). As these structures are part of a broader fan deposit, their radial spread in azimuth of these lineations are interpreted to be caused by frequent avulsion of a channel around a node found in sediment fans (Goudge et al., 2017).

The planform wedge shape of this unit and the almost radial array of raised lineations could be interpreted to be because of two reasons. Radial sediment fan shapes are often associated with relatively

preliminary stages of delta growth (Chamberlain et al., 2018). The Medium-light Toned Rough unit could be interpreted as an early-stage delta. However, radial shaped sediment fans are also distinctive of fluvial fan morphology (Weissmann et al., 2013). The orientation of the raised lineations are bimodal. This distribution could be caused by data bias from subsequent erosional conditions removing the central lineations. There have, however, been bimodal distributions of channel orientations found on Earth (e.g. Martinez et al., 2010) which are associated with palaeographic evolution. Subsequently, the bimodal distribution of the raised lineation orientations could also indicate that instead of one complete fan event, the Medium-light Toned Rough unit may be the result of reactivations with a change in preferred channel orientation over time.

Further inferences about the palaeoenvironment of the upper fan can be made from the geomorphology of the individual raised lineations. Assuming the raised lineations are inverted channels with only the channel fill preserved, these inverted channels would have an average 30:1 width to depth ratio and thus could be interpreted as sheets or thin channel belts rather than ribbons (e.g. Gibling, 2006). Sheets or thin channel belts at the distal portion of a sediment fan are often associated with fluvial fans (Weissmann et al., 2013). However, these channels are at the limits of detectability at the resolution of the HiRISE DTMs. Moreover, these inverted channels are only preserved remnants and may record only a portion of the channel's original thickness and width.

These remnant channels can be plotted back to the avulsion node (Figure 4.17) which helps make further inferences about the palaeoenvironment in which the fan was deposited in. The expansional apex of the upper fan unit is pinned to the channel-basin transition. This expansional apex is again interpreted to be the avulsion node which is supported by the average bearing of the preserved lineations plotting back to the point where the channel transitions into the lobe-shaped sedimentary body. This is a topographically pinned node on top of a delta sequence. Backstepping of avulsion nodes channelward do occur in deltas which can occur due to prior lobes raising the elevation of a channel (Chamberlain and Walling, 2019) or through rising shoreline levels (Chamberlain et al., 2018; Shaw and Mohrig, 2014). However, as the avulsion node of the upper fan unit is controlled by the channel-basin transition (or by the topography), this is more likely to be a fluvial fan as these are also characterised by their avulsion nodes being topographically controlled (Ganti et al., 2014; Prasajo et al., 2022).

Further evidence can be put forward to support a fluvial fan hypothesis. The gradual decrease in protrusions across the fan surface could be indicative of a gradual change in properties across the fan proposed as either a change in the abundance of boulders or properties of the rock which the fan is composed of (further discussed in section 2.2.1.2 under the Medium-light Toned Rough unit). A gradual change in properties across the fan surface is another key-characteristic of a fluvial fan (Weissmann et al., 2013).

Due to the nature of the channels, the channel orientations, the change in properties from the lower fan, the gradational change in properties across the upper fan surface, and the topographically pinned expansional apex, the upper fan is likely a fluvial fan. Although not recorded elsewhere on Mars yet,

fluvial fans superimposed on top of deltas have been observed on Earth, such as on the Colorado river delta when a dam dried out the delta and attempts to flood this delta ended up with formations of fluvial fans on top of the delta deposits due to rapid deposition in the 1930s (Sykes, 1937) and again in 1942 (Mueller et al., 2017). Therefore it is not unprecedented for a fluvial fan to be superimposed onto a delta when a dried out delta has been reactivated through flooding. Based on the geomorphology, a fan emplacement model for the Oxia Planum sediment fan is modelled (Figure 4.20).

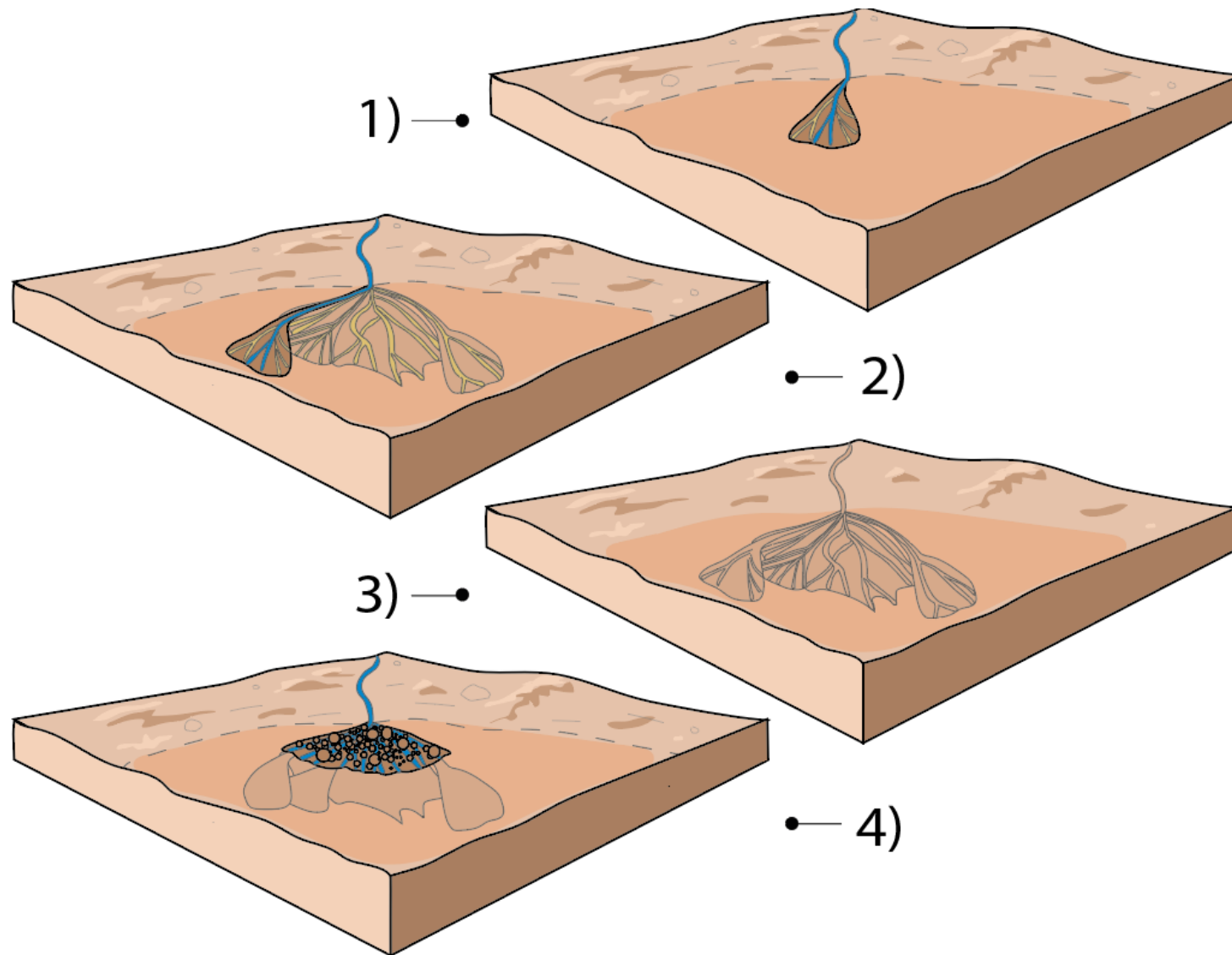


Figure 4.20: The fan emplacement model proposed for Oxia Planum. 1) Shows the formation of a sediment lobe over a flood plain or transient shallow lake. 2) Shows the evolution of the fan into a mature, unconfined fluvial distributary system or delta with sustained water and sediment flow over time. 3) Represents an extended dry period. 4) Shows the reactivation of the sediment fan as a relatively short-term depositional event under potentially higher flow rates.

4.6 Conclusions

The sediment fan in Oxia Planum is the youngest preserved unit in the study area. It overlies the phyllosilicate-bearing basement and the Pitted unit, a regional capping unit, but is cross-cut by a ridge associated with regional faulting, placing the sediment fan in stratigraphic context. The fan can be divided into two photogeologically distinct deposits: the Medium-light Toned Smooth unit and the Medium-light Toned Rough unit. The Medium-light Toned Smooth unit forms the lower fan, comprising six lobe- and ribbon-shaped bodies with cross-cutting relationships that record lateral and prograding avulsions, consistent with prolonged fluvial deposition. It also preserves candidate cross-bedding and planar layering. In contrast, the Medium-light Toned Rough unit forms the upper fan and is expressed as a single wedge-shaped body with radial inverted channels and a gradual change in sedimentary textures across its surface basinward – likely caused by changes to the cementation or grain-size of the rock, all features typical of a short-lived fluvial fan. An embedded and infilled crater between the two units indicates a prolonged hiatus in deposition, likely caused by more arid conditions. Whether the lower deposits represent a delta prograding into a shallow standing body of water or a fluvial fan terminating on a plain cannot be resolved from orbital data, but either interpretation has important implications for the scale and persistence of surface water in the Oxia Planum basin. With the Rosalind Franklin rover scheduled to land at Oxia Planum in 2028, the photogeologic map developed here provides context for testing these hypotheses and for supporting the mission’s goal of investigating past habitability.

Chapter 5: Synthesis, conclusions, and future work

5.1 Introduction

In this chapter, I consolidate the main findings from Chapters 2 through 4, evaluate how these results contribute to a broader understanding of regional climate evolution during the Late Noachian and Hesperian, as well as their implications for global Martian climate evolution. In addition, I critically assess the limitations associated with orbital datasets and in-situ rover investigations. Finally, I summarise the research of Chapters 2 through 4 and outline directions for future research, including terrestrial analogue studies, global comparative investigations, and future exploration missions on Mars and Titan.

5.2 Records of ancient climate change on mars

5.2.1 The Layered Sulfate unit at Gale crater

Chapters 2 and 3 provide the first ground-truth observations of the orbitally defined Layered Sulfate unit (LSu), prompting a re-evaluation of Gale crater's palaeoclimate during the late-Noachian and early-Hesperian. Prior to the Curiosity rover's in-situ exploration of the LSu, hypotheses for the local climate during the LSu's deposition were made from orbit and farfield observations. From orbit, a transition from clay-bearing strata into the layered sulfate-bearing strata (e.g. Fraeman et al., 2016; Milliken et al., 2010; Thomson et al., 2011) had long been hypothesised to record an aridification of Gale crater's paleoclimate, possibly representative of global events at the Noachian-Hesperian boundary (e.g. Bibring et al., 2006; Thomson et al., 2011). Hydrological modelling of Gale crater by Horvath and Andrews-Hanna (2021) similarly supported a secular drying trend. Additionally, Curiosity's earlier traverse through the clay-bearing and the spectrally mixed clay/sulfate-bearing strata (e.g. Sheppard et al., 2021a) recorded a succession of overall increasingly arid depositional environments (e.g. Edgar et al., 2024; Grotzinger et al., 2014; Gupta et al., 2022; Meyer et al., 2025), also appearing to support an aridification of Gale crater's palaeoclimate. As Curiosity commenced in-situ exploration of the LSu in August 2022, Chapters 2 and 3 therefore used the first ground-truth observations from the Curiosity rover to reconstruct the depositional environments recorded in the basal section of the LSu. Here, these results will be used to reassess the palaeoclimate of Gale crater during the late Noachian and early Hesperian to see whether they support previous aridification models.

5.2.1.1 How have the studies on the layered sulfate unit changed our understanding of the evolution of gale crater's palaeoclimate?

The results from Chapters 2 and 3 show that the LSu does not follow the overall aridification trend observed through the clay-bearing stratigraphy and the spectrally mixed clay/sulfate bearing

stratigraphy. As described in Sections 2.2 and 3.2, the clay-bearing strata record pervasive lacustrine, fluvial, and deltaic environments (Edgar et al., 2020; Grotzinger et al., 2014; Rivera-Hernández et al., 2019; Stack et al., 2019) and the spectrally mixed clay/sulfate bearing strata record a transition into more arid conditions, with metre-scale aeolian dunes interbedded with sporadic aqueous lenses (Edgar et al., 2024; Gupta et al., 2022; Meyer et al., 2025). While Chapters 2 and 3 show that the LSu record aeolian depositional environments with metre-scale aeolian dune cross-bedding (Section 3.4.1), tabular beds of aeolian sandsheet deposits (Sections 3.4.2 and 3.4.4), and aeolian scour-and-fill structures (Chapter 2), the presence of a fluvial lens (Section 2.5.4) and a laterally extensive lacustrine deposit (Section 3.4.3; Gupta et al., 2022; Mondro et al., 2025a; Mondro et al., 2025b; Weitz et al., 2023) indicate that surface water persisted beyond the mixed clay/sulfate-bearing strata. Notably, as discussed in Section 3.5.1.3, the aeolian sandsheet deposits were likely controlled by a near-surface water table, and, as discussed in Section 3.5.1.5, repetitive transitions between the light- and dark-toned planar-bedded sandsheet lithofacies were likely controlled by episodic intersections between the capillary fringe of the water-table and the depositional surface. Thus, both the mixed clay/sulfate bearing strata and the LSu are aeolian systems influenced by water. The architectural styles of both units, however, reveal distinct depositional settings. The mixed clay/sulfate-bearing strata are consistent with a dominantly ‘dry aeolian system’ with short-lived flooding of interdunes, whereas the layered sulfate-bearing strata reflect a ‘wet aeolian system’ shaped by a fluctuating water table (e.g. Mountney and Thompson, 2002). Consequently, while the layered sulfates may have formed under more arid conditions than the clay-bearing strata, they also represent a ‘wetter’ aeolian setting than the preceding mixed clay/sulfate bearing strata. Thus, the depositional environments recorded in the layered sulfates contradict a model of monotonic aridification during Mount Sharp group deposition.

Instead, results from Chapters 2 and 3 could be used to argue that the clay-sulfate transition record an overall period of climate warming and increasing humidity in the late Noachian to early Hesperian. As discussed in Section 3.5.1.3, a slowly rising water table was likely the primary control on both the accumulation and preservation of the ~60 m thick succession of aeolian sandsheet deposits in the LSu. Harris et al. (2024b) also identified candidate sandsheet deposits several hundred metres above the stratigraphy in Marker Band valley, suggesting that water-table rise may have continued to influence layered sulfate deposition beyond the section explored by *Curiosity*. While the magnitude of this rise is uncertain, the interpretation aligns with orbital studies showing that Gale crater, along with other crater basins, likely intersected a deep groundwater table that fluctuated regionally or even globally by ~800 m during the Hesperian (Horvath and Andrews-Hanna, 2017; Michalski and Bleacher, 2013; Salese et al., 2019). It is therefore plausible that the accumulation of the layered sulfates in Marker Band valley corresponded to a regional or even global upwelling of the water table on Mars. As aridity index and annual precipitation are the primary controls on groundwater level (Horvath and Andrews-Hanna, 2017), a rising water table may imply that the layered sulfates record deposition during a period of overall climate warming and increasing humidity, contrary to prior models of aridification during the

accumulation of the layered sulfate-bearing stratigraphy in the late-Noachian to early-Hesperian.

Results from Chapters 2 and 3 also demonstrate that the climate may have undergone small-scale fluctuations between more arid and humid conditions during the accumulation of the LSu in the late-Noachian to early-Hesperian. Super deflation surfaces at the base of the layered sulfate succession in Marker Band valley (Section 3.4.1) and the Marker Bed (Section 3.4.3; also see Rapin et al., 2021), likely formed from aeolian deflation during a period of water-table fall as discussed in Sections 3.4.1 and 3.4.3. Similarly, as discussed in Section 3.5.1.5 and Chapter 2, the aeolian scour-and-fill structures may also have been controlled by increased aeolian deflation during periods of water table fall. However, these fluctuations in climate conditions were likely small in scale, during a period of overall water-table rise in an increasingly humid climate at some point during the late-Noachian through to the early-Hesperian.

5.2.1.2 Can the sedimentary facies and architecture of the layered sulfates help constrain the source of the sulfate minerals?

A central question in discussions of the clay-sulfate transition in Gale crater concerns the origin of the sulfate minerals: did they form authigenically during deposition, or are they predominantly the product of later diagenetic alteration? Results from Chapter 3 suggest several sedimentary structures may point to a potential authigenic source of the sulfates. Adhesion structures (Section 3.4.4.4; Section 3.4.4.6; and potentially 3.4.4.2) indicate that the capillary fringe of a near-surface water table intermittently intersected the ground surface. Candidate lenticular crystal forms (Section 3.4.4.4) could indicate syndepositional crystal growth likely through evaporation at or near the surface when the sediment was deposited. As discussed in Section 3.4.1.1, the indurated pinstripes within the subcritically climbing translent strata likewise suggest preferential early cementation, potentially from groundwater evaporation in the subsurface (Fryberger and Schenk, 1988). In arid, warm depositional environments, groundwater evaporation often outpaces meteoric recharge, and can promote mineral precipitation as a cement (Worden and Burley, 2003), a process commonly documented in the subsurface of modern terrestrial sandsheet deposits (Basilici and Dal'Bó, 2014; Kocurek and Nielson, 1986). Sulfate-rich cements are not uncommon in arid areas from the evaporation of saline waters within the zone of groundwater capillary fringe rise (e.g. Pye, 1983). Thus, the sedimentary structures recorded in Chapter 3 suggest that the capillary fringe of the water-table intermittently intersected the surface, likely underwent evaporation leading to precipitation of cements, that could have been a potential source of the sulfate minerals.

Results from Chapters 2 and 3 suggest a diagenetic source for the sulfates is also likely. Figures 2.5c, 3.7e, and 3.10d-e document extensive overprinting by nodules of varying sizes and by cross-cutting veins, consistent with extensive diagenesis. Extensive diagenesis from post-depositional migration of brines through Gale crater's sedimentary rocks has been well documented throughout MSL's traverse leading to nodular overprinting and cross-cutting veins (Bristow et al., 2021b; Hurowitz

et al., 2017; Meyer et al., 2025). Sulfates have also been found to be concentrated in individual diagenetic nodules and a correlation between quantitative nodule abundance and bulk rock enrichment of sulfate has also been observed along Curiosity's earlier traverse of the spectrally mixed clay/sulfate stratigraphy (Seeger and Grotzinger, 2024). It is therefore likely that the diagenetic nodules observed in the overlying LSu may record a similar trend. As the sedimentary structures recorded in Chapter 3 indicate that the LSu's sulfates could be authigenic, diagenetic, or from a combination of both processes, further mineralogical and geochemical analyses will be required to constrain the origin of the sulfates.

5.2.1.3 Broader implications for layered sulfate stratigraphy and clay-sulfate transitions across Mars

The depositional environments recorded in the LSu are similar to ground-based rover observations of the layered sulfate stratigraphy in Meridiani Planum (Figure 1). At Meridiani Planum, Opportunity investigated ~7 m of sulfate-bearing strata (Edgar et al., 2014; Grotzinger et al., 2005; Herkenhoff et al., 2008; Metz et al., 2009b). Similar to the basal section of the LSu in Gale crater (Figure 1b, Section 3.4.1), the Burns formation also records metre-scale, cross-stratified sandstones from aeolian dunes (Figure 1a; Grotzinger et al., 2005; Kite and Conway, 2024) and tabular beds of aeolian sandsheet deposits (Edgar et al., 2014; Grotzinger et al., 2005; Metz et al., 2009b). In both formations, the sandsheet deposits are laterally extensive, traceable for hundreds of metres in rover panoramas, suggesting regionally significant sandsheet development (Chapter 3; Grotzinger et al., 2005; Metz et al., 2009b). In both cases, the conditions which led to the promotion of aeolian sandsheets and inhibition of dune formation were also likely damp surface conditions and early cementation (Chapter 3; Grotzinger et al., 2005; Metz et al., 2009b). The similar facies associations despite the ~7,000 km distance between the two sites also suggest that conditions which favoured the development of extensive aeolian sandsheets may not have been unique across the Martian surface.

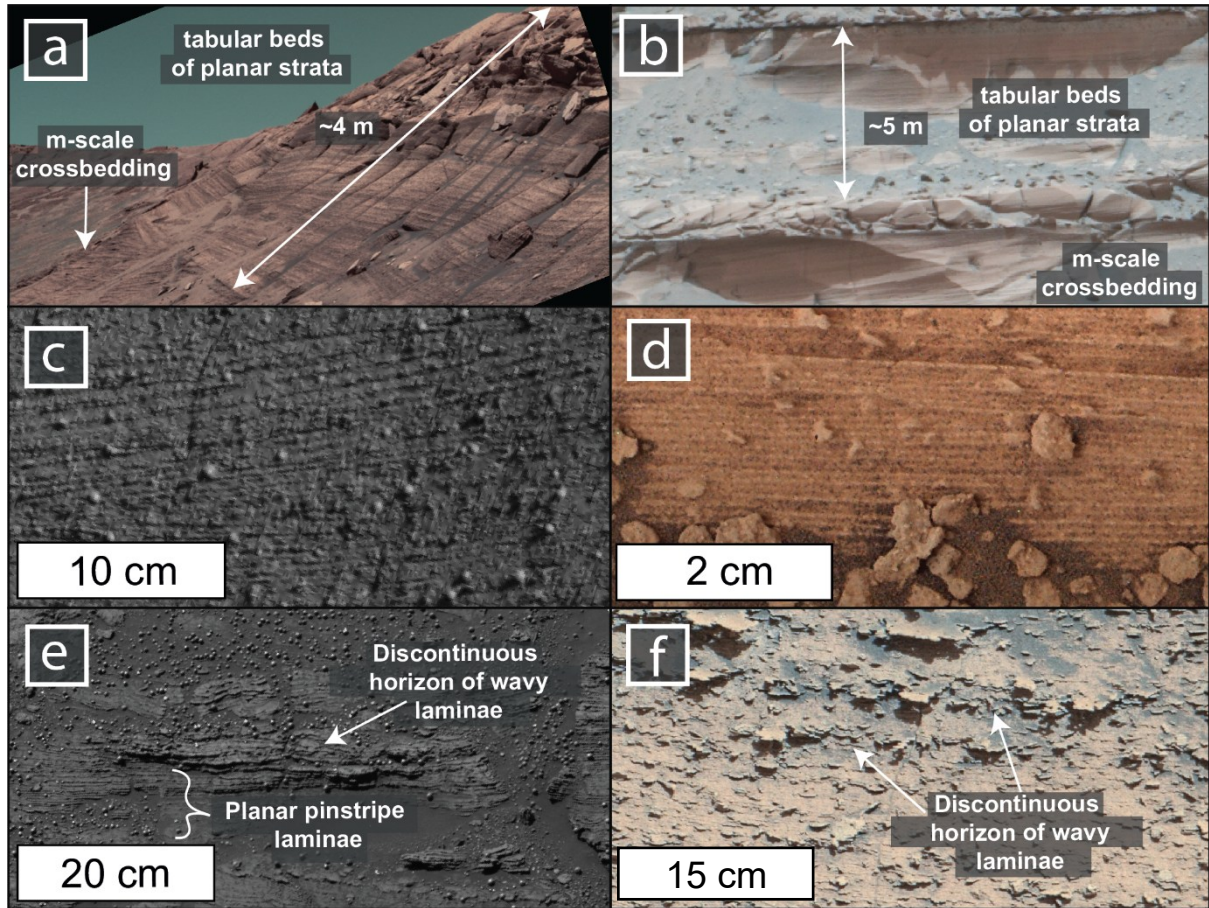


Figure 5.1: Comparisons between select facies characteristic of the aeolian sandsheets observed in Meridiani Planum and Gale crater. (a) Meridiani Planum aeolian sandsheet deposits within the Burns formation based on Grotzinger et al. (2005). Note that the aeolian sandsheet deposits here only comprise a ~4 m thick bed and aeolian sandsheet deposit thicknesses rarely exceed ~5 m elsewhere in Meridiani Planum (e.g. Metz et al., 2009b). (Opportunity Pancam mosaic PL_p2440 acquired on Sol 278, at Sol 277 end-of-drive location, Endurance crater). (b) Gale crater LSu aeolian sandsheet deposits that make up the Catrmani member of the Mirador formation. (Mastcam mosaic MR_mcam103144 acquired on Sol 3662, at Sol 3658 end-of-drive location, Gale crater). Note that both deposits truncate beds of m-scale, aeolian cross-bedding. (c) Well-developed pinstripe texture which has likely formed due to differential cementation of translent strata from Grotzinger et al. (2005) (Opportunity Pancam mosaic PR_p2557 acquired on Sol 307, at Sol 303 end-of-drive location, Endurance crater). (d) Well-developed pinstripe texture which has likely formed due to differential cementation of translent strata from Chapter 3 (Kastria Spring target, 25 cm stand off, Gale crater). Note the difference in laminae thickness between (c) and (d) still fall in the expected range of laminae thicknesses for subcritically climbing translent strata (e.g. Chakraborty and Chakraborty, 2001; Fryberger et al., 1979; Fryberger and Schenk, 1988; Hunter, 1977; Kocurek and Nielson, 1986). (e) Corrugated and irregularly wavy laminae suggested to indicate possible damp sandsheet conditions in Meridiani Planum taken from Grotzinger et al. (2005) (Opportunity Pancam Mosaic PR_p2561, acquired on Sol 045, at Sol 043 end-of-drive location, Eagle crater). (f) Discontinuous horizons of corrugated laminae interpreted as adhesion structures see Chapter 3 Figure 3.15 for more detailed annotations (Mastcam mosaic MR_mcam103987 acquired on Sol 3812, at Sol 3810 end-of-drive location, Gale crater). Image credits: (a, c, e) NASA/JPL-Caltech/Cornell/ASU; (b, d, f) NASA/JPL-Caltech/MSSS.

However, the different topographic settings of the two localities have important implications for the local climate. Whilst the aeolian sandsheet deposits in Gale crater likely accumulated in association with a near-surface water table (Section 3.5.1), and Grotzinger et al. (2005) similarly attributed the aeolian sandsheets at Meridiani Planum to near-surface groundwater, Gale crater is a deep topographic basin that would have intersected a slowly rising groundwater table (e.g. Horvath and Andrews-Hanna, 2021; Salese et al., 2019). In contrast, Meridiani Planum sits ~3000 m higher in elevation, at the top of a regional plateau, and groundwater upwelling to the surface at this location would require substantial rise of the regional water table and prolonged periods of climate warming (e.g. Andrews-Hanna et al., 2010). The presence of rip-up clasts and possible festoon cross-bedding within the Meridiani sandsheets provides evidence for unchannelised flooding of the aeolian surface (Grotzinger et al., 2005; Metz et al., 2009b). Damp and cementing conditions needed to inhibit dune formation and promote sandsheet accumulation can also result from periodic surface flooding rather than groundwater rise (e.g. Kocurek and Nielson, 1986). Therefore, periodic run-off and surface flooding may also have contributed to the accumulation of aeolian sandsheets at Meridiani Planum, also acknowledged by Grotzinger *et al.* (2005).

As the clay-sulfate transition at Gale crater is the first to be explored in-situ by a rover, the research presented in Chapters 2 and 3 have implications for clay-sulfate transitions observed elsewhere on Mars. As described in Chapter 1, clay-sulfate transitions are widespread across the Martian surface and have previously been hypothesised to have resulted from a shift to a more arid climate during the Noachian-Hesperian boundary (e.g. Bibring et al., 2006; Ehlmann and Edwards, 2014; McLennan et al., 2019; Milliken et al., 2010). Although the results from Chapter 3 demonstrate that there may be some correlation with depositional environments promoting drying surface conditions leading to the precipitation of cements in the subsurface (e.g. Section 5.2.1.2), the orbitally defined clay-sulfate transition region at Gale crater was not deposited during a period of a drying climate and instead potentially during a climate of increasing humidity but set within an overall aridification trend (Section 5.2.1.1). Consequently, transitions from clay to sulfate mineralogy across other successions on Mars may also reflect local depositional environments or later diagenetic modification instead of accumulation under the aridification of the regional climate.

Results from Chapter 3 may also have implications for layered stratigraphy elsewhere on Mars. Laterally extensive deposits elsewhere on Mars can exhibit decimetre- to metre-scale planar layering and reach total thicknesses of over a few kilometres (e.g. Cadieux and Kah, 2015; Edgett and Sarkar, 2021; Grindrod and Warner, 2014; Grotzinger and Milliken, 2012; Quinn and Ehlmann, 2019; Weitz et al., 2010), appearing similar to the layered sulfates at Gale crater from orbit (Figure 2; Chapter 3; Fraeman et al., 2016; Sheppard et al., 2021a; Weitz et al., 2010). These deposits have been attributed to a wide range of potential mechanisms from lava flows to lakes to dust falls (e.g. Cadieux and Kah, 2015; Edgett and Sarkar, 2021; Grotzinger and Milliken, 2012; Quinn and Ehlmann, 2019). Chapter 3 shows how thick aeolian sandsheet successions can also accumulate hundreds-of-metres-thick, planar-stratified successions on Mars. Therefore, accumulations of aeolian sandsheets is a plausible

emplacement mechanism for the extensive layered sedimentary deposits across Mars, which formed in an era of general aridity, but with some limited influence of surface and/or groundwater.

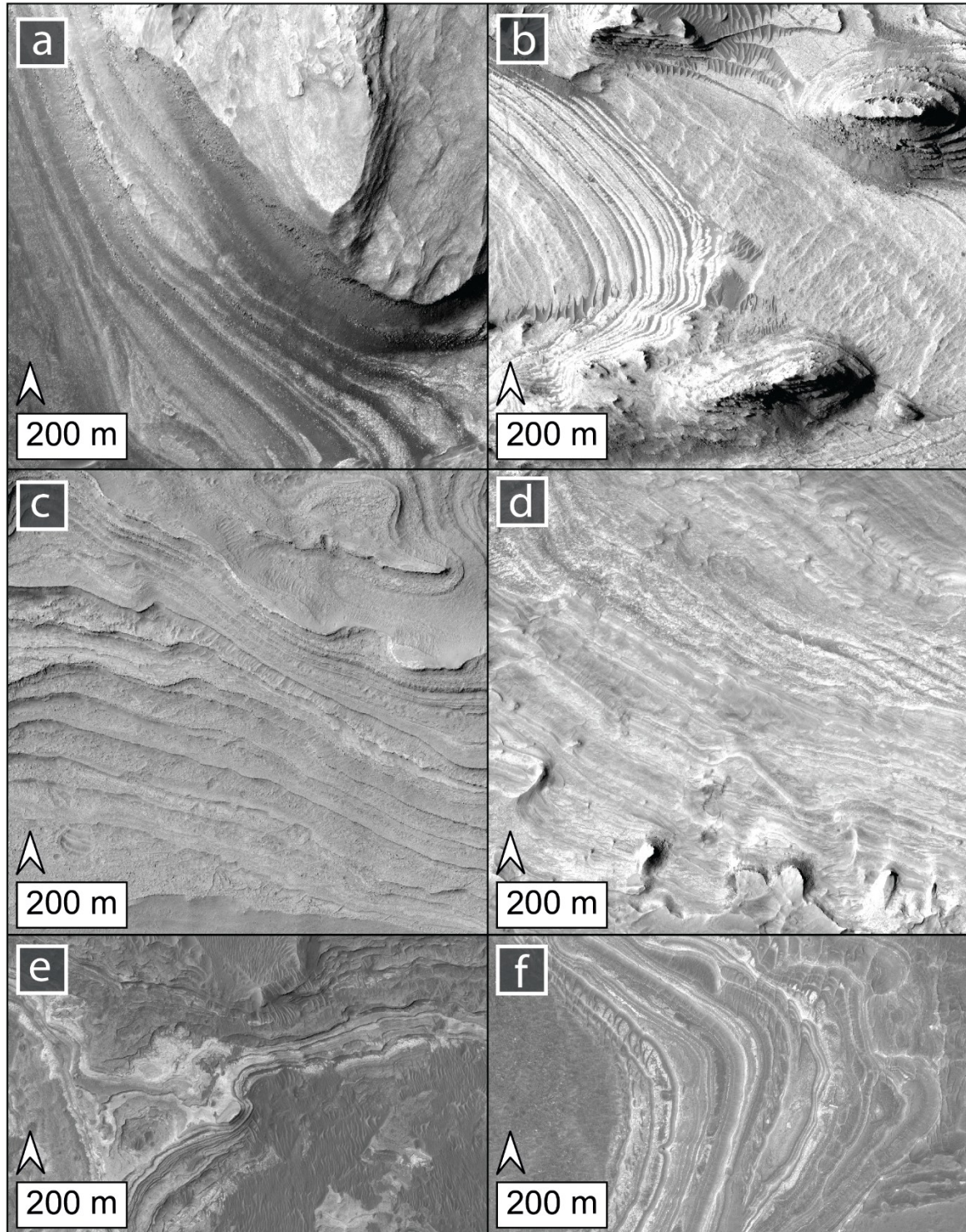


Figure 5.2: LSu in Gale crater compared to selected other layered deposits on Mars adapted from Weitz et al. (2010). (a) Gale crater. HiRISE image PSP_007501_1750. (b) Becquerel crater. HiRISE image PSP_001918_1735. (c) Melas Chasma. HiRISE image PSP_007667_1700. (d) Meridiani Planum. HiRISE image PSP_007440_1845. (e) Ladon Vallis. HiRISE image PSP_006637_1590. (f) ILD of Juventae Chasma. HiRISE image PSP_003579_1755.

5.2.2 The Sediment Fan at Oxia Planum

Chapter 4 directly addresses the research questions introduced in Chapter 1 concerning the depositional history of the sediment fan in Oxia Planum and the implications for the regional climate evolution during the late Noachian to early Hesperian. The sediment fan in Oxia Planum occurs at the terminus of the Coogoon Vallis system at the margin of the Oxia Planum basin. Prior to my investigation detailed in Chapter 4, Fawdon et al. (2022) identified at least two possible periods of fluvial activity in the Coogoon Vallis system: paleo-coogoon vallis and Neo-coogoon vallis. Paleo-Coogoon vallis can be linked to the formation of fluvial sinuous ridges and possible floodplains linked to the blue-toned, clay-bearing unit that formed in the mid-Noachian or prior (Davis et al., 2023). Reactivation of Coogoon Vallis occurred as Neo-Coogoon vallis, where the sediment fan occurs at the terminus at an unknown period prior to the formation of the Kilkhampston crater (Fawdon et al., 2022) in the early Hesperian (Grindrod et al., 2022). As Coogoon vallis extends for hundreds of kilometres with a considerable catchment area of $\sim 2.1 \times 10^5 \text{ km}^2$ (Fawdon et al., 2022), one of the central aims of Chapter 4 was to then constrain the depositional history of the sediment fan and what this might mean for the nature of water activity and the regional climate during the Late Noachian through to the Early Hesperian.

5.2.2.1 How has the investigation into the sediment fan in Chapter 4 changed our understanding of Oxia Planum's climate evolution?

Chapter 4 shows that fluvial activity in Coogoon Vallis was episodic and varied in duration. As discussed in Section 4.6.1, the Oxia Planum sediment fan post-dates the clay-bearing unit linked to the formation of Palaeo-Coogoon Vallis (e.g. Davis et al., 2023; Fawdon et al., 2022). The fan itself comprises two superimposed photogeologic units with distinct morphologies. An infilled crater between the two units implies a significant time gap of tens of millions to several tens of millions of years (Quantin-Nataf et al., 2021) where the sediment fan was inactive, indicating at least two episodes of fluvial activity during the emplacement of Neo-Coogoon vallis. So, in total, there were at least three periods where Coogoon Vallis was active. Morphological evidence also offers insight into their relative longevity. The lower fan, formed of widely spaced, curvilinear ribbon- and lobe-shaped bodies with cross-cutting relationships, resembles fans formed under sustained, low-sediment discharge over extended periods of time (Hartley et al., 2010). Comparable fans on Earth can take thousands to tens of thousands of years to form (Donselaar et al., 2022), suggesting the lower fan may have been a similarly long-lived system, consistent with inferred ages for nearby fluvial systems in Arabia Terra that also formed in the mid-Noachian to early Hesperian (Davis et al., 2019). In contrast, the upper fan, with its wedge-shaped form and radial straight channels, resembles fans formed rapidly under high sediment supply and discharge (e.g. Hartley et al., 2010), or during early-stage development (e.g. Chamberlain et al., 2018), both implying shorter time-scales for the formation of the upper fan. Superimposed catastrophic flood deposits on fluvial fans on Earth can also have similar morphologies (Gao et al.,

2020; McCarthy and Ellery, 1998; Moscariello, 2018; Wolski and Murray-Hudson, 2006). Therefore, Chapter 4 demonstrates that Coogoon Vallis experienced at least three discrete episodes of fluvial activity, each with differing durations.

The episodic fluvial activity at Coogoon Vallis can be used to infer multiple episodes of climate warming during the mid-Noachian to early Hesperian. As some global climate models of Mars for this period generally fail to produce mean annual temperatures above freezing (e.g. Forget and Leconte, 2014; Wordsworth et al., 2013; Wordsworth et al., 2015), regional, or possibly global, warming events are required to explain the availability of surface water for fluvial activity in Coogoon Vallis. The longevity of fluvial activity inferred for Palaeo-Coogoon Vallis (Davis et al., 2023) and the lower sediment fan suggests that earlier warming episodes during the mid-Noachian to Early-Hesperian were relatively long-lived to support persistent precipitation or snow-melt and runoff (e.g. Craddock and Howard, 2002; Hoke and Hynek, 2009; Hynek et al., 2010). In contrast, the third phase of fluvial activity that formed the upper sediment fan was likely relatively short-lived. Thus, this event may have resulted from a relatively brief period of warming. It could also be postulated that fluvial activity was attributed to localised meltwater input rather than widespread precipitation (e.g. Fastook and Head, 2015; Wordsworth et al., 2013; Wordsworth et al., 2015). Chapter 4 therefore indicates that the Oxia Planum region experienced at least three warming events, with earlier episodes more prolonged than later ones.

Another aim of Chapter 4 (as described in Chapter 1) was to reconstruct the depositional environments that deposited the sediment fan. Whilst Chapter 4 was able to provide a history of emplacement of the fan, I could not confidently resolve the depositional environments at the orbital resolution. Prior to the investigation of Chapter 4, there were two end member hypotheses on whether the sediment fan in Oxia Planum was a river debouching into a standing body of water to form a delta (e.g. Quantin-Nataf et al., 2021) or a fluvial/alluvial fan debuting over an unconfined, dry alluvial plain (e.g. Ivanov et al., 2020). Chapter 4 found that the morphology of the sediment fan in Oxia Planum, comprised of seven ribbon and lobe-shaped sediment fan bodies that record lateral, prograde, retrograde avulsions, and cross-cutting relationships, was similar to both deltas terminating in a standing body of water (e.g. Bhattacharya, 2006; Chamberlain et al., 2018; Ganti et al., 2014; Reynolds, 2022) and a fluvial fan terminating over an unconfined dry alluvial plain (e.g. Cain and Mountney, 2009; Donselaar et al., 2022; Hartley et al., 2010; Weissmann et al., 2013). Either scenario would have important implications for the nature of water and sediment in the basin. If the sediment fan was a delta, it would have entered a standing body of water that likely stretched through the entire basin and possibly even open to the Northern lowlands (Quantin-Nataf et al., 2021). If the sediment fan was an alluvial fan, then water and sediment from Neo-Coogoon vallis was likely only constrained to the sediment fan and may not have continued further into the basin due downstream losses in discharge (e.g. Cain and Mountney, 2009; Donselaar et al., 2022; Hartley et al., 2010; Weissmann et al., 2013). To further resolve the nature of water and sediment in the basin, diagnostic sedimentary textures and facies of deltaic environments and architectural features such as clinoforms would need to be identified which cannot be resolved from

orbit (see Section 5.5.2.3).

5.2.2.2 Do other sediment fans across Mars exhibit changes in depositional style during fan reactivation, and does this have any implications for climate evolution?

Although no studies have yet undertaken global investigations focused on fan reactivation on Mars, several well-studied sediment fans exhibit at least two stages of deposition under changing depositional environments, similar to the results for the sediment fan in Oxia Planum from Chapter 4 (Figure 5.3; Table 5.1). These sediment fans either have a lower fan deposit characterised by either a branched fan deposit (Goudge et al., 2017; Morgan et al., 2022; Chapter 4; Pondrelli et al., 2011; Stack et al., 2020b) or a wedge of radially arranged sinuous and cross-cutting, inverted channels (Grindrod et al., 2018; Williams et al., 2011). These earlier deposits consistently display smoother photogeologic textures than their late-stage counterparts and show clearly defined layering in cross-section (Table 5.1). In contrast, the final stage of deposition exhibits a distinctly different photogeology, characterised in all cases by a coarse-textured, unstructured appearance at metre-scale resolution. It remains difficult to determine whether this texture is inherent to the deposits or the result of weathering from less consolidated material, but the presence of embedded boulders in scarps at Coprates Catena (Grindrod et al., 2018) and Holden crater (Grant et al., 2008), along with ground-truth observations of conglomerates at Jezero crater by Perseverance (Figure 4; Caravaca et al., 2024; Mangold et al., 2024; Mangold et al., 2021), confirms a coarse sedimentary texture in at least some late-stage deposits. The morphology of these late-stage deposits also differ from the earlier fan deposits, falling broadly into wedge-shaped deposits (Chapter 4; Davis et al., 2018; Goudge et al., 2017; Grindrod et al., 2018; Stack et al., 2020b; Williams and Weitz, 2014) and more ‘swath’-shaped deposits (Pondrelli et al., 2011; Williams et al., 2011). The late-stage deposits often overly unconformities (Caravaca et al., 2024; Grant et al., 2008; Mangold et al., 2024) or show evidence for substantial hiatuses (Chapter 4), indicating that fan deposition was episodic.

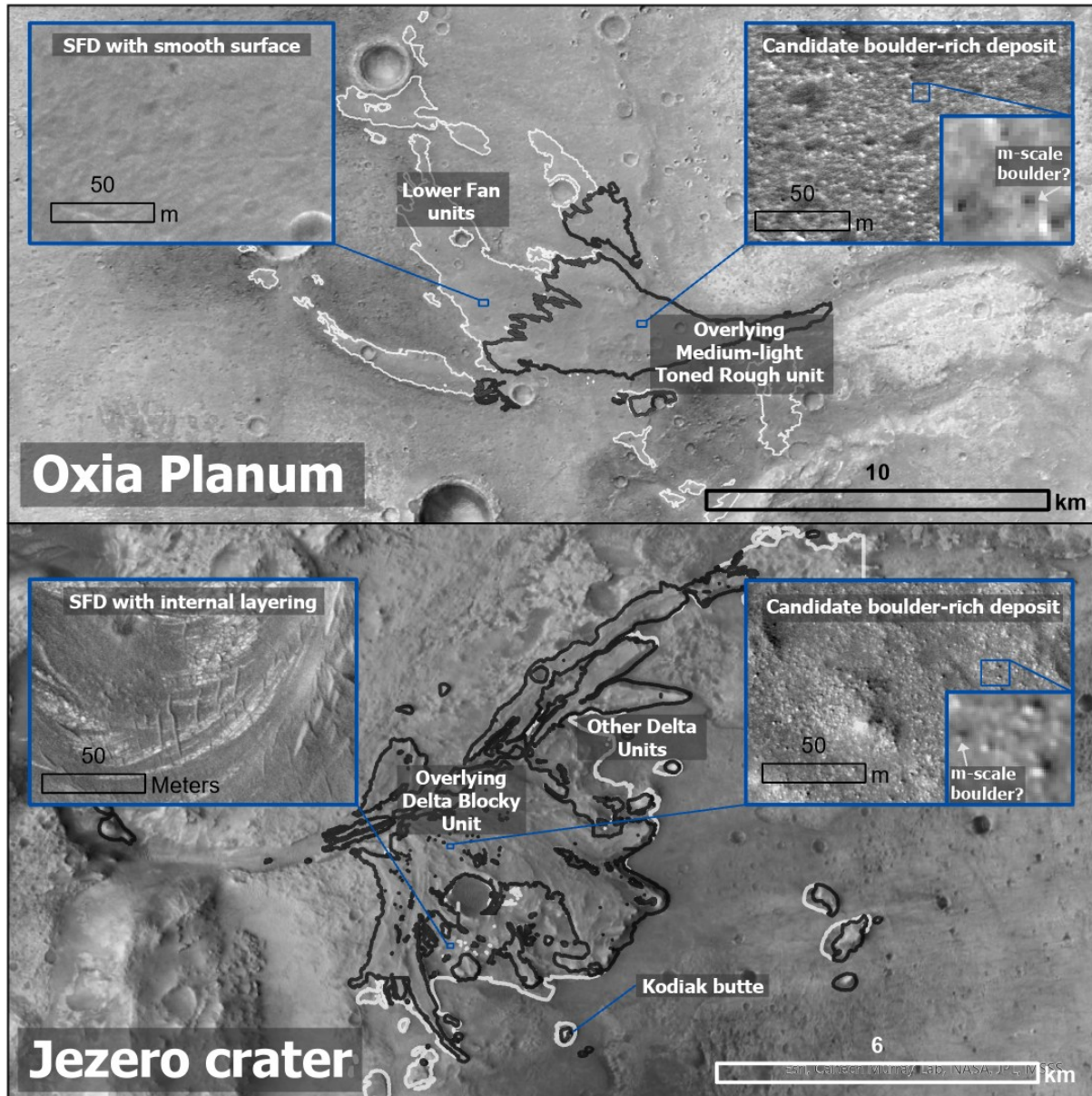


Figure 5.3: Map of the Oxa Planum fan (top panel) and Jezero fan (bottom panel), Sediment fan outlined in white and the late-stage fan, candidate boulder-rich deposit outlined in black (from Chapter 4; Stack et al., 2020b). Basemap is Context Camera (CTX; Malin et al., 2007). Image credits: ESRI/Caltech Murray Lab/NASA/JPL/MSSS. Inset panels show contrasting surface textures within the region: left shows a relatively smooth texture and layered textures from lower SFDs, while right shows a rough texture associated with candidate boulder-rich deposits with candidate m-scale boulders highlighted. Inset panel basemaps are from HiRISE (image seqids ESP_039721_1980_RED and PSP_002387_1985_RED).

These preliminary comparisons of the results from Chapter 4 to several other multi-staged sediment fans across Mars have implications for the Martian climate. First, the early stages of fan emplacement often exhibited distinct layering within individual channels and preserved multiple cross-cutting relationships (e.g. Grindrod et al., 2018; Pondrelli et al., 2011; Stack et al., 2020b; Williams et al., 2011), signifying stacked channel deposits formed through vertical aggradation consistent with sustained fluvial activity rather than a single surface runoff event (e.g. Davis et al., 2019; Chapter 4). Similar to Chapter 4, the longevity of these systems would require a long-lived period of climate

warming to support persistent precipitation or snow-melt and runoff (e.g. Craddock and Howard, 2002; Hoke and Hynek, 2009; Hynek et al., 2010). For late-stage fan emplacement, m-scale blocky textures were often interpreted as evidence for boulder-rich conglomerates from high-energy flood deposits (e.g. Goudge et al., 2017; Grindrod et al., 2018; Pondrelli et al., 2011). While difficult to constrain from orbit, rover observations at Jezero crater have identified conglomerates that sharply incise the underlying deltaic sandstones with no obvious sorting or bedding (Figure 4; Caravaca et al., 2024; Mangold et al., 2024). These conglomerates have been linked to a widespread boulder-rich deposit identified on the Jezero fan from orbit (Figure 5.3; Goudge et al., 2018; Stack et al., 2020b) and attributed to late-stage floods disconnected from lake activity (Caravaca et al., 2024; Mangold et al., 2024). In Coprates Catena, Grindrod et al. (2018) documented possible inverse grading, a feature commonly also associated with flash-flood events on Mars. The morphologies of the late-stage fan deposits are also comparable to those produced by high-discharge events in terrestrial fluvial and alluvial fans, albeit across a spectrum of sediment supply conditions (Hartley et al., 2010). If some of these late-stage deposits were from flash floods, then these could have been triggered by glacial surges, heavy seasonal snowmelts, or arid climate flash floods from storms (e.g. Mangold et al., 2024). Collectively, these systems record episodic activity during the Late Noachian and Hesperian, and all may preserve evidence for late-stage, high-energy flooding. Further research is needed to better constrain the depositional environments, shared surface processes, and possible climate triggers responsible for these processes (see section 5.5.2).

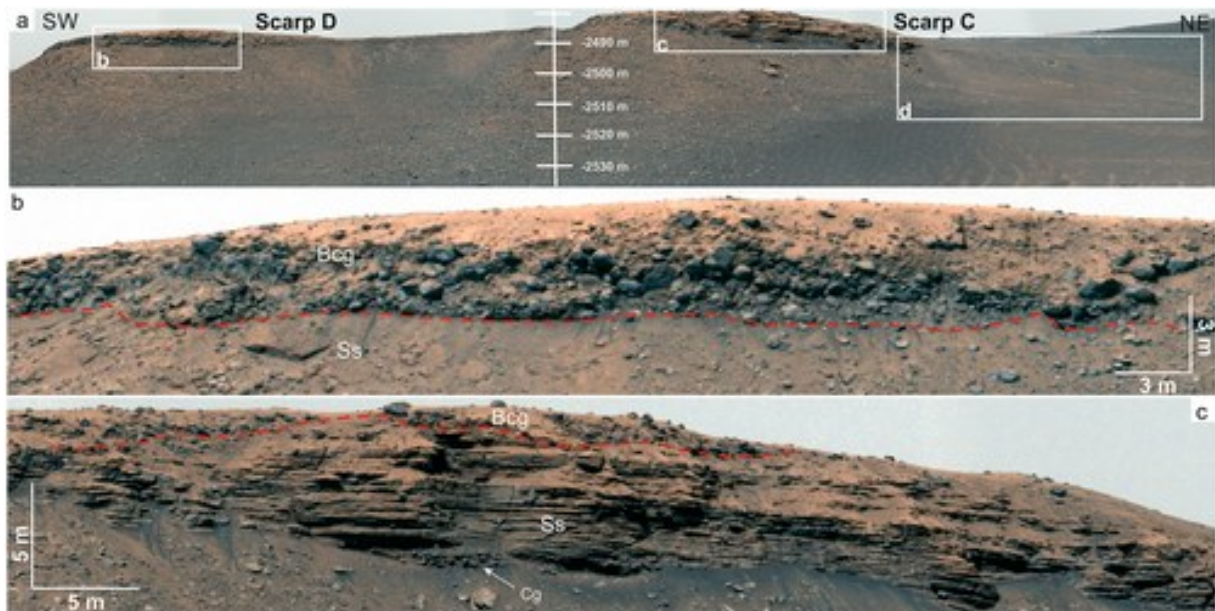


Figure 5.4: Supercam example of conglomerates at Jezero crater overlying a sharp, irregular, erosional surface that may be an unconformity taken from Mangold et al. (2024).

Location	Lower fan: Unit morphology and photogeology	Upper fan: Unit morphology and photogeology	Interpretation	Age	Reference
Oxia Planum	Geomorphology: Fan deposit comprised of multiple ribbon and lobe-shaped bodies. Photogeology: Comprised of a comparatively smooth unit that exhibits dark- and light-toned layering.	Geomorphology: Fan deposit comprised of a wedge-shaped deposit terminating as radial inverted ridges. Photogeology: Comprised of a medium-light toned rough unit characterised by m-scale blocks and boulders.	Lower Fan: Long-lived fluvial fan or delta. Upper Fan: Short-lived, late-stage fluvial fan deposit, possibly related to a flooding event.	Mid-Noachian through Early Hesperian	Chapter 4
Jezero crater	Geomorphology: Fan deposit comprised of multiple ribbon and lobe-shaped bodies. Photogeology: Comprised of multiple units that are comparatively smooth and exhibits dark- and light-toned layering.	Geomorphology: Fan deposit extends across the fan surface as a wedge-shape. Maintains ridges organised into overlapping triangular deposits. Overlies unconformity detected from in-situ rover exploration (Figure 4; Mangold et al., 2024) Photogeology: Rough, unstructured deposit comprised of m-scale boulders.	Lower Fan: Delta Upper Fan: Late-stage, catastrophic flood-deposit (Mangold et al., 2024)	Noachian to Hesperian	Goudge et al. (2017) Stack et al. (2020b) Mangold et al. (2021) Mangold et al. (2024) Caravaca et al. (2024) Stack et al. (2024) Grant et al. (2008)
Holden crater	Geomorphology: not visible Photogeology: Layered dark- and light-toned layers.	Geomorphology: Unconformably overlies the lower fan deposit. Photogeology: m-scale blocky/boulder rich texture.	Lower unit: Deposition in a distal alluvial fan and/or lacustrine environment. Upper unit: High energy and short-lived flood event.	Late Noachian	
Coprates Catena (Fan 1 and Fan 2)	Geomorphology: Radially extending ridges with a digitate appearance interstratified in recessively weathering massive strata. Ridges occasionally cross-cut each other.	Geomorphology: Wedge-shaped deposits comprised of multiple flat, stacked benches. Photogeology: Comprised of alternating coarser sedimentary texture comprised of m-scale boulders.	Lower fan deposit: Fluvial or alluvial fans terminating as a delta in a shallow lake. Upper fan deposit: Later period of deposition as an alluvial fan.	Somewhere between the Noachian-Hesperian boundary and the Early Amazonian	Grindrod et al. (2018)

	Photogeology: Ridged material interfingering with sheetlike, layered deposits.	Cross-sections record possible inverse grading.			
Western Terra Tyrrhena	Geomorphology: Two lobes. Surfaces characterised by abundant radial linear ridges that occasional cross-cut each other. Photogeology: Comparatively smooth texture. Layering visible on cross-sections and inverted channels.	Geomorphology: A wide swath that extends basinward over the lower fan surface as a linear path. Photogeology: Rugged texture comprised of m-scale blocks/boulders [way of saying no organisation].	Lower fan deposit: Deposition as an alluvial fan. Upper fan deposit: Transition into late-stage debris flows suggested to represent a decline in available water and/or change in sediment supply.	Late Noachian	Williams et al. (2011)
Eberswalde crater	Geomorphology: Fan deposit comprised of multiple ribbon and lobe-shaped bodies. Surface is comprised of abundant sinuous channels with preserved meander belts. Photogeology: Relatively smooth. Light- and dark-toned layering. Light-layers on fan surface contain m-scale rough textures.	Geomorphology: The youngest lobe E, sits on top of the fan as a low-sinuosity inverted, digitate ridge that terminates as a lobe-shaped body. Overlies a significant unconformity. Photogeology: Does not exhibit any layering. Comprised of a rough, m-scale blocky texture.	Lower Fan deposit: Delta over a shallow body of water. Upper Fan deposit: Terminal fan from a high-energy flooding event.	Late Noachian	Pondrelli et al. (2011)
South-west Melas basin (Eastern fan and Western fan)	Geomorphology: Cross-cutting inverted, sinuous bifurcating channels and lobate deposits with finger-like terminations. Photogeology: Limited descriptions.	Geomorphology: Wedge-shaped deposit with inverted channels radiating from a central point, channels preserved on their top surfaces Photogeology: Limited descriptions.	Lower Fan deposits: Delta over a shallow body of water. Upper Fan deposits: Late-stage alluvial fan or delta-top.		Williams and Weitz (2014) Davis et al. (2018)

Table 5.1: Summary of Martian sediment fans that record at least two distinct stages of deposition. For each locality, the morphology, photogeology, relative geologic age, and interpretations for the lower and upper fan deposits are given.

5.2.2 What do Oxia Planum and Gale crater tell us about possible climates during the mid-Noachian to early-Hesperian

Two different areas of Mars, one representing a surface environment (Oxia Planum) and the other a deep basin setting (Gale crater), record evidence for episodes of climate warming during the mid-Noachian to early-Hesperian. At Oxia Planum, the lower fan records prolonged fluvial activity, likely spanning tens to hundreds of thousands of years, sustained by overland flow from the Neo-Coogoon Vallis system. These timescales are consistent with other long-lived fluvial systems documented in Arabia Terra (Balme et al., 2020; Davis et al., 2023; Davis et al., 2019; Fawdon et al., 2022). In a deep basin setting, the thick successions of sandsheets in Gale crater's layered sulfate deposits indicate interaction with a rising, near-surface water table, possibly linked to a regionally extensive or even global water table (e.g. Salese et al., 2019). Together, these records suggest episodes of climate warming and increased humidity during the mid-Noachian to early-Hesperian, leading to sustained runoff on the surface and a rising water-table that influenced sediment accumulation in deep basins.

These warming events were also likely episodic. Multiple reactivations of Coogoon Vallis and the Oxia Planum sediment fan, followed by prolonged inactivity, and the onset of more fluvial activity. Unconformities within the layered sulfates also record significant hiatuses in deposition that may be linked to water table fall and increased aeolian deflation. These observations from two contrasting settings suggest that climate conditions during the mid-Noachian to early-Hesperian, the climate oscillated between warmer conditions capable of sustaining long-lived fluvial systems and deep but rising water-tables and drier periods marked by prolonged periods of aridity.

5.3 Challenges in Understanding Climate Change on Mars: Image Resolution and Gaps in the Rock Record

A persistent limitation in reconstructing the Martian palaeoclimate is that datasets are often at resolutions too low to resolve the fine-scale sedimentary features needed to distinguish between similar-appearing depositional environments. Results from Chapters 2 and 3 demonstrate how previous interpretations from lower data resolutions can lead to hypotheses for vastly different depositional environments than the ones identified in the higher resolution datasets from in-situ exploration by the rover. For example, at limited data-resolutions available prior to in-situ exploration of the layered sulfates, Rapin et al. (2021) interpreted broad scours interbedded with planar strata in the layered sulfates as fluvial channels and associated overbank deposits. When Curiosity commenced in-situ exploration of the layered sulfates however, no fluvial infill was identified with one exception (Section 2.5.4). Instead, the higher resolution datasets from Mastcam and MAHLI led to the identification of mm-scale pinstripe laminae (Figure 1) in the laterally equivalent light-toned, planar layered material that infilled the scours (Chapters 2 and 3), placing the broad, u-shaped scours from localised aeolian

deflation instead of fluvial erosion (Chapter 2). Likewise, the Marker Bed, described in Section 3.4.3, was initially interpreted from orbit as a volcanic ash or lag deposit (Weitz et al., 2022). But it was only with ground-based detection of cm-scale symmetric ripples (Section 5.4.3; Gupta et al., 2023b; Mondro et al., 2025a; Mondro et al., 2025b; Weitz et al., 2023) that provided the first evidence of an ice-free shallow lake open to the atmosphere on Mars. This same limitation was encountered in Chapter 4, where the resolution of orbital data alone could not help constrain whether the Oxia Planum sediment fan was a delta or fluvial fan without additional information on grain-scale textures or architectural indicators such as clinoforms, as both systems can appear similar from orbit. However, orbital datasets provide a key exploratory framework for guiding time-limited exploration by rovers.

In-situ exploration from landed missions also have their own limitations as a rover's spatial coverage of a region is inherently limited. Curiosity's traverse crosses only a small fraction of Gale crater, meaning that it is difficult to constrain whether the deposits encountered represented basin-wide conditions or if they were completely localised to that small part of the traverse. Whilst Chapter 3's geologic mapping demonstrates the aeolian sandsheet units can be laterally traced for hundreds of metres (Figures 2.4 and 3.1), on Earth, aeolian sandsheets are often spatially restricted to erg margins or also are adjacent to other environments like sabkhas, salt pans, or lakeshores (e.g. Kocurek, 1996; Kocurek and Nielson, 1986; Mountney, 2006a; Mountney, 2012). It is therefore plausible that the aeolian sandsheets laterally transitioned into a large erg or salt pan in Gale crater but lies beyond the rover's coverage, which would have implications for the region's palaeoclimate. Orbital surveys can address some of these limitations by placing observations into a wider geological context. As shown in Chapter 4, orbital datasets can help resolve the timing and persistence of fluvial activity in a given region. This thesis demonstrates that integrating orbital and in-situ datasets (e.g. Chapters 2, 3, and 5) provides the most effective approach for reconstructing Mars's geologic history prior to human exploration.

Another limitation in reconstructing the palaeoclimate from either orbit or in-situ exploration by landed missions is the limited preservation state of sedimentary deposits. Chapters 2 and 3 demonstrate how many sedimentary structures encountered by rover exploration are significantly overprinted by diagenesis (Figure 2.5), exhumed by subsequent aeolian erosion (Section 2.5.3), and mantled by dust and other surficial deposits leading to challenges in reconstructing past depositional environments. This was especially true for the reconstruction on how the deepening shallow lake deposits transitioned into the overlying aeolian sandsheet deposits (Section 3.4.4.1) as the contact was so overprinted by diagenesis that fine-scale sedimentary facies could not be resolved.

5.4 Conclusions

Chapter 2 uses ground-based rover observations to document decametre-wide, concave-up scour-and-fill structures in the Chenapau member of the Mirador formation at the base of Gale crater's Layered Sulfate-bearing unit. Detailed facies and architectural analysis reveals two types of scours that

formed under different mechanisms. A small scour with lateral pinching and a dark-toned appearing infill comprised of cm-scale cross-bedding (Class B) was consistent with fluvial scouring and infill. In contrast, broad and symmetrical concave scours reaching up to ~60 m in width with a draping light-toned infill continuous with the surrounding sandsheet stratigraphy (Class A), suggests an alternate aeolian in origin. More specifically, the geometry, architecture, and infill facies are consistent with blowout structures. The concentration of aeolian scour-and-fill structures in a single stratigraphic interval suggests a transition into an environment with intense wind activity during the early development of Gale crater's Layered Sulfate-bearing unit, consistent with a long-term trend of aridifying *palaeoenvironments* during the late-Noachian and early-Hesperian.

Chapter 3 investigates the broader stratigraphic context of the basal section of the LSu exposed in buttes around Marker Band valley and demonstrates that simple aeolian dunes and aeolian sandsheet deposition remained the dominant mode of sedimentation throughout this succession, supporting the long-term trend of aridifying *depositional palaeoenvironments*. However, aeolian sandsheets were likely promoted by damp surface conditions inhibiting dune development, likely a near-surface water table. Furthermore, stratigraphic variations between light- and dark-toned planar lithofacies and the aeolian palaeo-scours point to short-lived fluctuations in water table, sediment supply, and wind strength, with darker, adhesion-structure-bearing units representing intermittent intersection of the water-table and the surface. Isolated fluvial and lacustrine deposits in the Amapari and Chenapau members provide further evidence that intermittent surface water was periodically present. The substantial (~60 m) thickness of aeolian sandsheets far exceeds typical terrestrial analogues and likely reflects gradual overall water-table rise, contradicting the interpretation of an aridifying *climate* trajectory during the accumulation of the LSu and may instead point to a period of increasing climate humidity to explain the rise in water table during the late-Noachian and early-Hesperian.

The HiRISE-scale mapping of the sediment fan at Oxia Planum in Chapter 4 demonstrates that other regions of Mars were also warm enough to sustain surface water and that activity was episodic. The sediment fan comprises two photogeologic units with distinct geomorphologies. The lower fan is comprised of a Medium-light Toned Smooth unit with cratered surfaces, polygonal fractures and planar-layering organised and is organised into overlapping lobe- and ribbon-shaped bodies that record repeated lateral and basinward avulsion cycles. From HiRISE resolutions, it cannot be resolved whether the lower fan is a fluvial fan or a delta but the multi-branched morphology is consistent with a long-lived period of fan activity. In contrast, the upper fan is comprised of the Medium-light Toned Rough unit characterised by a m-scale blocky texture that is most pronounced near the channel apex and decreases basinward. The upper fan is comprised of a single wedge-shaped fan body with radial inverted channels and limited bifurcation, consistent with high discharge rates and a comparatively shorter period of fan formation. These superimposed, distinct fan systems have climate implications for Oxia Planum, suggesting multiple warming events during the mid-Noachian to early-Hesperian, with the first capable of sustaining prolonged precipitation or snowmelt, and a final period that was comparatively

brief and potentially driven by localised meltwater input.

Both Oxia Planum and Gale crater record evidence for episodes of climate warming and increased humidity during the mid-Noachian to early-Hesperian, resulting in prolonged fluvial runoff at Oxia Planum and a rising water table in Gale crater. These warming events were likely episodic, as shown by multiple reactivations of the sediment fan at Oxia Planum and by super-deflation surfaces that record hiatuses in sedimentation and increased aeolian deflation during the accumulation of Gale crater's layered sulfate deposits, all within the broader context of an overall aridification trend.

5.5 Future Work

5.5.1 Investigating Spatial and Temporal Drivers of Dry Accumulation, Damp and Cementing Conditions, and Deflation in Aeolian Sandsheets

Chapters 2 and 3 of this thesis have shown that there are still a number of unknowns in regards to understanding depositional and erosional environments in aeolian sandsheets. Section 3.5.1.5 highlights the need for further research on how spatial variations - and temporal variations - in surface processes operating on an aeolian sandsheet can impact sandsheet dynamics and accumulation, particularly where damp sandsheets, dry sandsheets, and deflation coexist. These research gaps made it difficult to resolve whether climate fluctuations or internal dynamics of a sandsheet system led to the interbedded dry and damp sandsheets and deflation that make up the Marker Band valley layered sulfate succession. The extent to which vegetation influences sandsheet dynamics and controls the accumulation of aeolian sandsheets is also unknown. In addition, Chapter 2 also highlights the research gap in understanding blowout formation without the role of vegetation, as active blowouts have only been studied in vegetated aeolian systems (e.g. Hesp, 2002). As such, some terrestrial analogues may offer a means to investigate the accumulation of dynamic aeolian sandsheets in systems minimally influenced by vegetation.

Ideal terrestrial analogues for Martian aeolian sandsheets must balance several environmental and logistical considerations. Whilst large active sandsheets exist on Earth that record both damp and dry modes of deposition and deflation (Al-Hurban, 2014; Feder et al., 2018; Fryberger et al., 1983; Kocurek and Nielson, 1986; Koster, 1988; Mountney and Russell, 2004), and recent studies have documented thick sandsheet successions in the rock record comparable to those in Marker Band valley (Abrantes et al., 2020; Bállico et al., 2017; Basilici et al., 2021; Ferronatto et al., 2019; Murton et al., 2025; Simplicio and Basilici, 2015), suitable examples where vegetation plays a limited role are rare on Earth. Ideal analogue field sites to begin to test the questions posed in Chapters 2 and 3 must therefore feature sand sheets which accumulated in regions with sparse vegetation, damp conditions regulated by shallow water tables, localised aeolian deflation, and where field access is relatively safe.

5.5.1.1 Modern sandsheet analogues

Kuwait Sandsheets: One promising candidate for a Mars sandsheet analogue could be the Kuwait desert. Here, active sandsheets cover large portions of the country (e.g. Khalaf and Al-Ajmi, 1993). Prevailing drought conditions since the 1990s led to hypersaline groundwater prohibiting growth of new shrubs (Al-Hurban, 2014; Misak and Hussain, 2022), leading to an ideal analogue to study desert processes with only minimal impact of vegetation. Active sandsheets, damp sandsheets, blowouts, and sabkhas have been observed to coexist (Al-Hurban, 2014; Khalaf, 1989; Mountney, 2006a), thus an ideal location to test how spatial heterogeneity of surface processes in these systems can influence sandsheet dynamics and accumulation.

Possible research directions could involve both remote-sensing and field-based approaches. High resolution satellite coverage of Kuwait goes as far back as the early 1990s (e.g. Khalaf and Al-Ajmi, 1993), enabling the ability for GIS analysis to see how aeolian sandsheets have evolved spatially over time and how they responded to a changing climate. Complementary field investigations could include trenching the sandsheets to examine vertical facies relationships (e.g. Kocurek et al., 2007), ground-penetrating radar to characterise subsurface architecture (e.g. Jewell et al., 2017), and detailed sedimentological logging to reconstruct facies associations and their spatial transitions. Together, these methods would support a stratigraphic reconstruction of how dynamic sandsheets evolve through time and space in the absence of extensive vegetation.

White sands: Another promising site for future analogue studies is the White Sands dune field in New Mexico, USA. Previous research has identified interdune deposits that record dry aeolian sandsheets, spatially restricted damp conditions, and evidence of localised deflation (Ewing and Kocurek, 2010; Kocurek et al., 2007), similar to the environments described in Chapters 2 and 3. Figure 6 shows a strip of smooth terrain (>6 km long, and ~450 m wide) between ribbed terrain and the dune field identified from orbit that does not show much evidence of widespread vegetation. The etched terrain (Figure 5.5) is likely a deflation surface, exposing truncated dune cross-strata where cemented wind-ripples stand in relief against adjacent grain-flow deposits (e.g. Grotzinger et al., 2005; Kocurek et al., 2007). In contrast, the smoother strip (Figure 5.5) appears to mantle this deflated surface and likely marks the accumulation of an aeolian sandsheet at the erg margin (e.g. Mountney, 2006a). These candidate sandsheets display both light- and dark- toned patches (Figure 5.5). The darker areas likely mark areas where organics have grown in response to damp conditions (e.g. Kocurek et al., 2007) whilst the lighter patches likely represent dry deposition. Scattered pockmarks (Figure 5.5), 20 - 50 m in diameter, that occasionally expose patches of the ribbed terrain, may represent blowouts or remnant pits from parabolic dunes. The clear juxtaposition of the appearance of damp, dry, and deflated sandsheets in close proximity, together with decades of high-resolution imagery, makes White Sands an ideal site to explore how spatial variations in moisture availability, sediment transport, and wind regime shape sandsheet dynamics and how they develop over time.

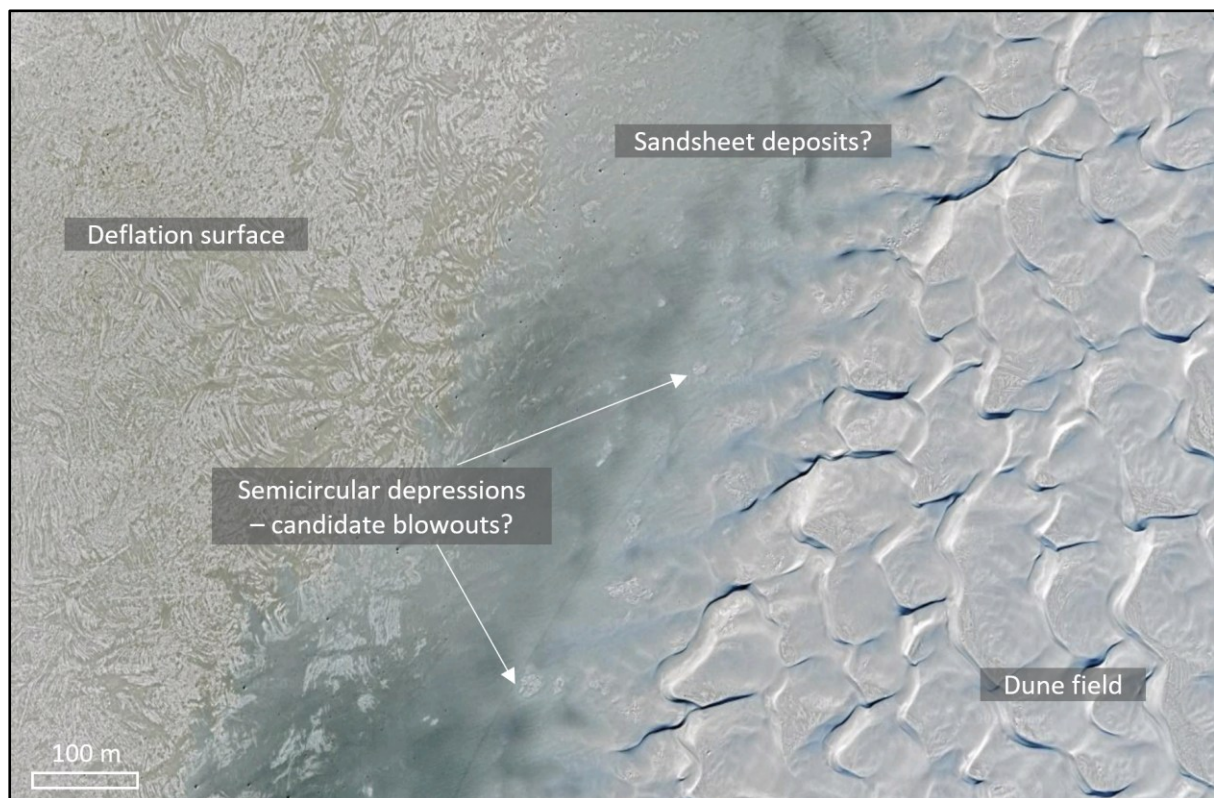


Figure 5.5: Satellite image of white sands erg margin including deflated plains, candidate aeolian sandsheets, and dune field. Note, the light-toned and dark-toned appearing patches in the potential sandsheet deposits linked to organic growth over damp sandsheet deposits (e.g. Kocurek et al., 2007). Semicircular depressions could represent candidate aeolian scouring from blowouts. Image credit: 'White Sands.' Google Earth, Google, Date of Access (15, July, 2025).

5.5.1.2 Ancient Sandsheet Analogues

Ancient terrestrial sandsheet successions also provide valuable Martian analogues. Thick sandsheet accumulations in the rock record, especially those formed before the widespread appearance of land plants (ca 420 Ma; Davies and Gibling, 2010; Rainbird, 1992), can serve as analogues for Martian sandsheets by capturing the different facies and surface processes that occur in the absence of vegetation. A number of Precambrian successions in Brazil, such as the Palaeoproterozoic Bandeirinha and Galho do Miguel formations, preserve thick stacked sandsheets, damp interdunes, salt flats, and scour-and-fill structures, closely resembling the facies variability seen in Marker Band valley (Abrantes et al., 2020; Mesquita et al., 2021; Simplicio and Basilici, 2015). A comparative review of these Precambrian sandsheet successions could help refine our understanding of sandsheet accumulation mechanisms where vegetation was absent and whether there are common genetic relationships between facies that may hint towards internal dynamics of an aeolian sandsheet system.

A second, related avenue is the investigation of blowouts in ancient, vegetation-free systems. The Lower Cretaceous São Sebastião Formation remains one of a few published examples where blowouts have been identified in the rock record (Ferronato et al., 2019). This rarity stands in stark contrast to their ubiquity in modern aeolian systems (e.g. Hesp, 2002), raising questions about their recognition, preservation potential, and diagnostic criteria in the rock record. In a Precambrian sandsheet succession, the Galho do Miguel formation, broad, shallow scour-and-fill structures with draping aeolian sandsheet infill have been identified and attributed to aeolian scour-and-fill structures (Abrantes et al., 2020), but their scale, architecture, and associated facies bear a striking resemblance to blowouts described in Chapter 2. A systematic review of possible blowout structures in the terrestrial rock record could determine whether such features have been overlooked. Future fieldwork could be conducted on some of these features, including the ones in the Galho de Miguel formation, to further constrain whether these are in fact blowouts and the genetic relationships with the surrounding facies to help reconstruct the depositional environments in which these formed.

5.5.2 Global Survey on Late-stage Fan Reactivations on Mars

As discussed in Section 5.2.2.2, several sediment fan systems across Mars are superimposed by deposits from late-stage reactivations of the fan system similar to the Oxia Planum sediment fan system described in Chapter 4 of this thesis. This raises broader questions: how common were late-stage reactivations of Martian fan systems, and what do these reactivations reveal about climate variability during the late Noachian and Hesperian? To answer these questions, a logical future research direction could be to undertake a global survey of Martian sediment fans to identify and catalogue superimposed deposits that post-date primary fan formation. Out of the ~1,676 Martian fan-shaped deposits identified to date (Morgan et al., 2022), between 605 to 869 sites are covered by high-resolution HiRISE images suitable for resolving metre-scale textures (Figure 5.8). Systematically reviewing these sediment fans

for potential superimposed, late-stage deposits may help establish how widespread these deposits were and allow for exploration of trends in their global distribution relative to the latitude, depositional environments, and timings of emplacement. More in-depth comparative analysis of a select few pristine fans along with the ground-truthed sediment fan at Jezero crater may further aid in understanding the environmental and climate drivers of late-stage fan reactivation.

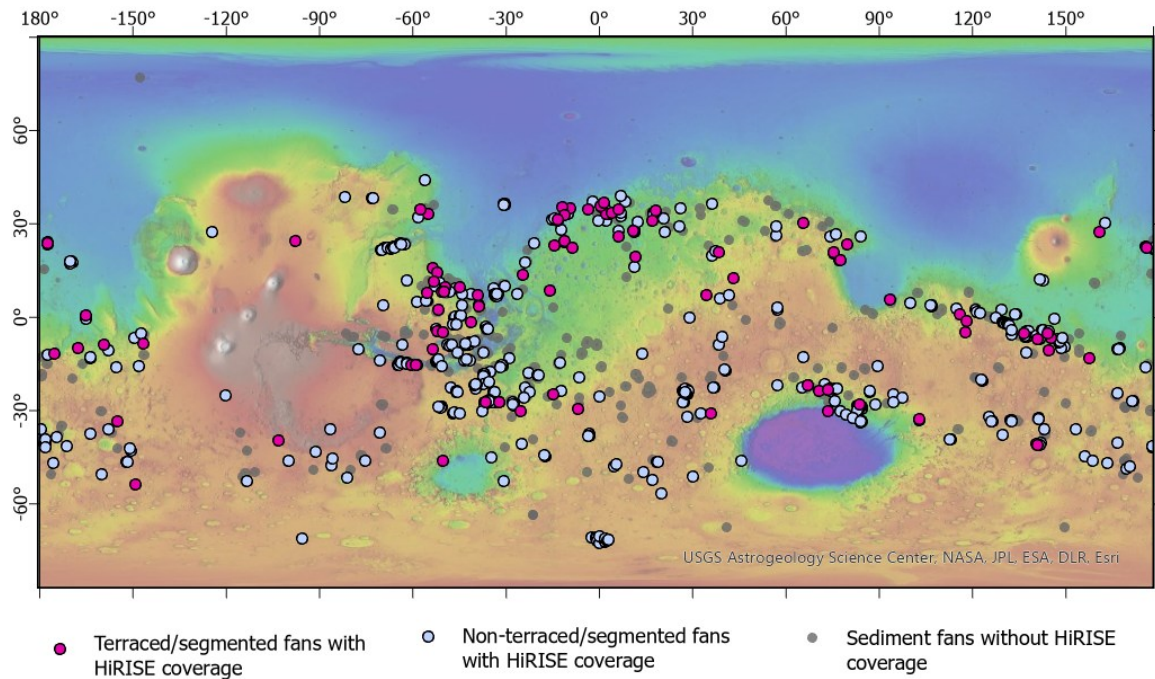


Figure 5.6: Global database of sediment fans across the Martian surface, categorized by morphology and HiRISE coverage. The 605 sediment fans with available HiRISE coverage subdivided into terraced/segmented fans (Mondro, 2022; Morgan et al., 2022) and non-terraced/segmented fans, while locations of fans lacking HiRISE coverage are also indicated. This map highlights both the morphological diversity of sediment fans and the extent of high-resolution imaging coverage of sediment fans across the surface of Mars. Basemap is HRSC-MOLA (Ferguson et al., 2018).

5.5.3 Testing Sedimentary Hypotheses with Upcoming Planetary Missions

Finally, with Curiosity still exploring the layered sulfates, Rosalind Franklin and Dragonfly set to launch in 2028 to investigate sedimentary systems on Mars and Saturn's moon Titan, and a number of other planetary missions with the goal of reconstructing past depositional systems, it is worthwhile discussing future research directions in regard to landed planetary missions.

5.5.3.1 Curiosity rover

Continued exploration of the layered sulfate-bearing stratigraphy by Curiosity at Gale crater offers an opportunity to test whether the aeolian depositional environments identified in Marker Band valley persisted further upslope. The valley exposes only the basal ~100 m of the layered sulfates, estimated from orbit to be 500 - 900 m thick (Fraeman et al., 2016). Far-field mosaics taken 500 - 2,500 m upslope reveal laterally extensive tabular beds of planar strata, broad scours, and possible cross-bedding that appear similar to far-field observations of Marker Band valley stratigraphy which led to the hypothesis

that aeolian dunes, sandsheets, and blowouts continued to accumulate up-section (Harris et al., 2024b). As Curiosity ascends Mount Sharp, this hypothesis can be tested with future nearfield observations of sedimentary facies and architecture to help determine whether wet aeolian conditions persisted, and how climate and water availability evolved over time.

5.5.3.2 Rosalind Franklin

The sediment fan described in Chapter 4 lies within one of the extended Rosalind Franklin landing ellipses, providing the opportunity to test how and when sediment and water were introduced into the Oxia Planum basin. Ground-truthing by the rover could help resolve competing interpretations on whether the fan formed as a delta in a standing body of water or as a fluvial fan terminating over dry terrain. A deltaic origin would imply that a substantial body of water once covered much of the landing ellipse and may even have connected to the northern lowlands (Quantin-Nataf et al., 2021), carrying significant implications for local and regional geology. Sedimentary structures such as clinoforms, characteristic of Gilbert-type deltas (Mangold et al., 2024; Mangold et al., 2021) could be identified with PanCam (Coates et al., 2017), while finer-scale sedimentary facies found in lacustrine environments - such as laminated mudstones (Stack et al., 2024) or symmetric ripples (Section 3.4.3; Gupta et al., 2023b; Mondro et al., 2025a; Mondro et al., 2025b; Weitz et al., 2023) - may be resolvable using both PanCam and the close-up imager (CLUPI; Josset et al., 2017). The photogeologic map developed in Chapter 4 could support future operations planning by helping to prioritise image targets and contact science if the rover were to land nearby.

5.5.3.3 Future helicopter missions to Titan and Valles Marineris

Valles Marineris, currently under active consideration for future aerial exploration platforms (e.g. Young et al., 2021; Zappek et al., 2023), presents a compelling target to investigate how sulfate-bearing sedimentary rocks formed elsewhere on Mars, and whether their depositional environments reflect the same climate-driven processes observed in Gale crater. The canyon walls of Valles Marineris expose kilometre-scale cliff-sections through layered sedimentary rock, many of which contain sulfate-bearing strata (Chojnacki and Hynek, 2008; Fueten et al., 2008; Roach et al., 2010; Wiseman et al., 2010). Previous orbital studies have hypothesised that the layers accumulated in lacustrine, volcanic, mass wasting, or aeolian depositional environments (Chojnacki and Hynek, 2008), but the depositional history remains unresolved. Aerial exploration platforms could provide the first detailed observations of bedding geometries, stratigraphic contacts, and sedimentary structures along vertical canyon exposures. The stratigraphic mapping approach developed in Chapters 2 and 3 of this thesis, which integrated periodic, targeted cliff-section mosaics and orbital images, could be adapted to document the sedimentary architecture of Valles Marineris and reconstruct its depositional history with future aerial exploration platforms. The sulfate-bearing stratigraphy here is of particular interest as it offers an opportunity to evaluate whether similar conditions to the ones that formed aeolian sandsheet deposits

in Gale crater (e.g. Chapter 3) and Meridiani Planum (Grotzinger et al., 2005), also occurred in Valles Marineris or whether the layered sulfates could have occurred under different depositional conditions.

It would also be valuable to investigate whether comparable depositional environments to those described in this thesis also occurred on other planetary bodies such as Titan, particularly as the Dragonfly mission prepares to explore its equatorial dune fields (Lorenz et al., 2018). Orbital data already reveals that a significant proportion of Titan's surface is covered in dunes and interdunes (Radebaugh, 2013). Depositional interdunes may have accumulated as dry or damp aeolian sandsheets (Barnes et al., 2008; Radebaugh et al., 2008) similar to the ones described in Chapter 3. Lineated plains have also been hypothesised as dune fields stabilized by cement or damp conditions (Radebaugh et al., 2011), an environment conducive to the formation of aeolian blowout features (Chapter 2). Titan also has a number of sediment fans which may show periodic precipitation over arid regions (Birch et al., 2016), similar to the intermittently active sediment fan at Oxia Planum discussed in Chapter 4.

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Appendices

Appendix I. Figures

I.1 Map of key image locations used for stratigraphic mapping the LSu.

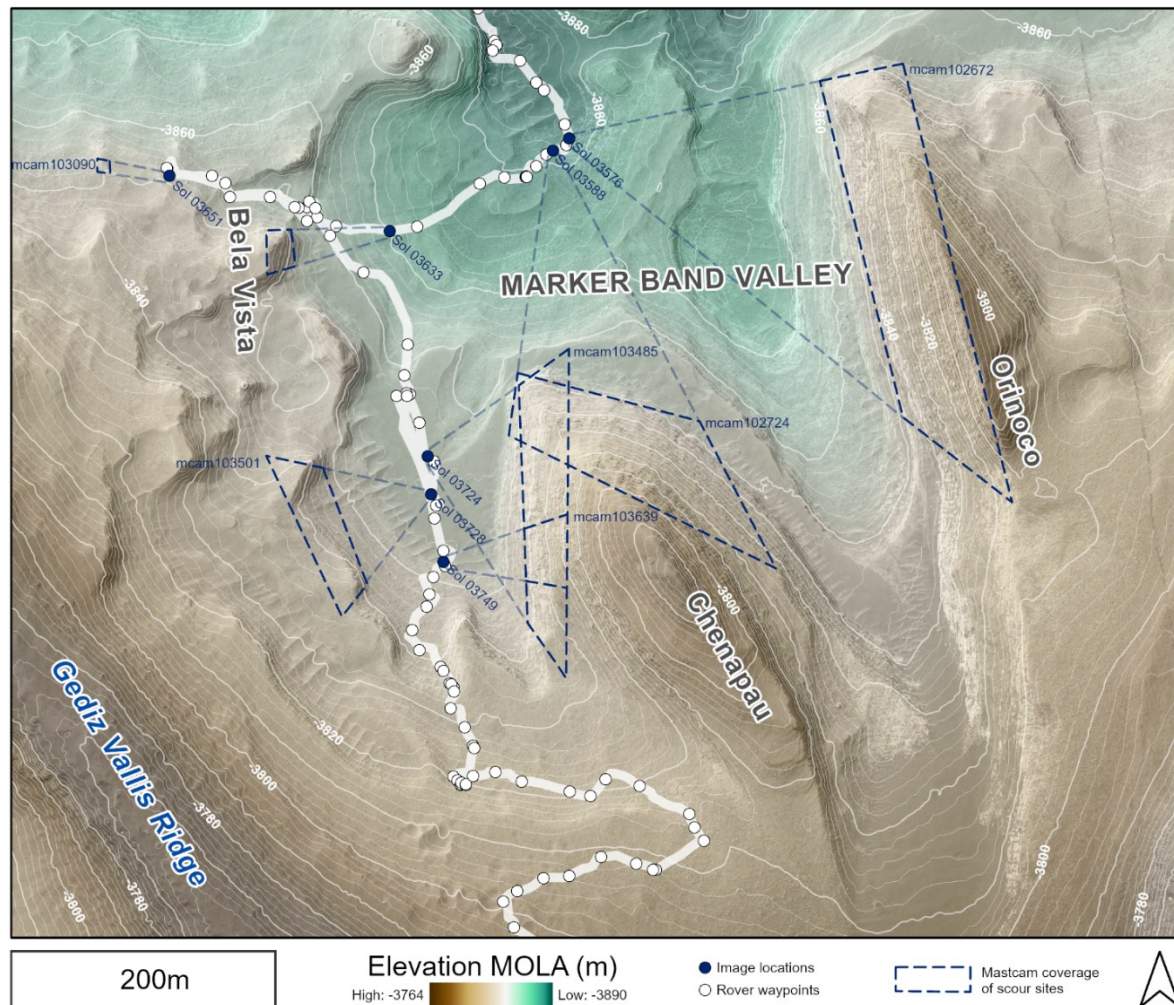


Figure I.1. Detailed map of the Marker Band valley region of the rover traverse depicting key imaging locations and Mastcam coverage of scour sites. High Resolution Imaging Science Experiment (HiRISE) mosaic (~25 cm per pixel; Calef and Parker, 2016) overlain by a Mars Orbital Laser Altimeter (MOLA) and HiRISE derived digital elevation model (~1 m per pixel resolution; Calef and Parker, 2016; Smith et al., 2001). Image credits: NASA/JPL/University of Arizona.

I.2 Map of stratigraphic contacts compared to the scour-and-fill structures.

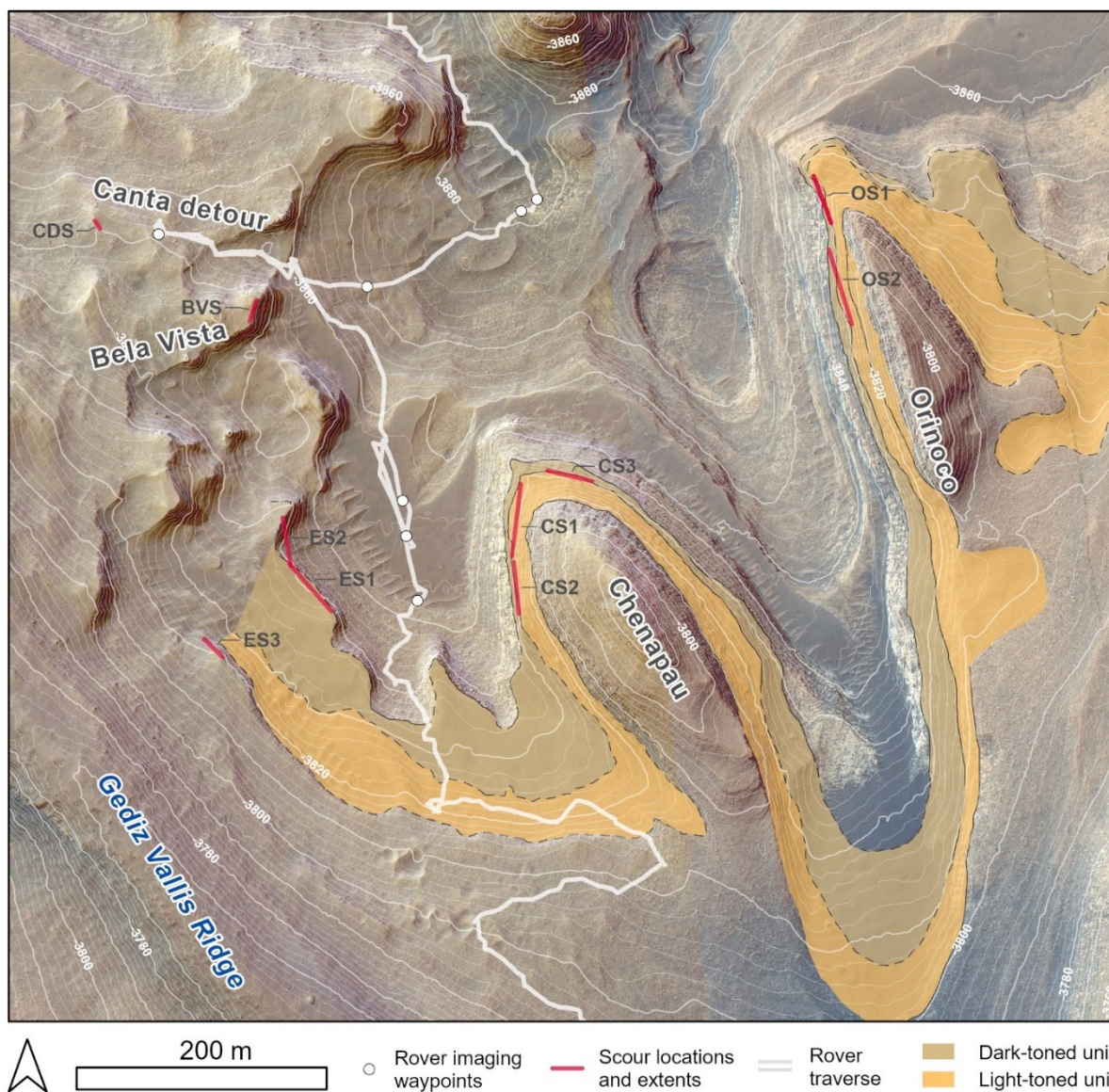


Figure I.2. Map of the stratigraphic contacts of the light- and dark-toned units. HiRISE mosaic overlain by a MOLA and HiRISE derived hillshade shows the Marker Band valley region of the rover traverse (from sols 3576 to 3749). Solid lines delineate measured contacts while dashed lines delineate inferred contacts. Red lines show scour locations, orientations, and extents with shorthand labels. White line shows Curiosity's traverse. Black-rimmed circles show the key imaging locations. Image credits: NASA/JPL/University of Arizona.

I.3. Example for manual measurements of apparent laminae thickness measurements.



Figure I.3.1. Locations and summary of the apparent laminae thickness measurements done by hand used for the light-toned unit along a transect orthogonal to the laminae (Kastria Spring target, 5 cm stand off). Note: measurements were taken manually to ensure greater consistency when comparing laminae thickness across section. Image credits: NASA/JPL-Caltech/MSSS.

I.4. Measurements of apparent laminae thickness using peak detection on greyscale pixel intensities.

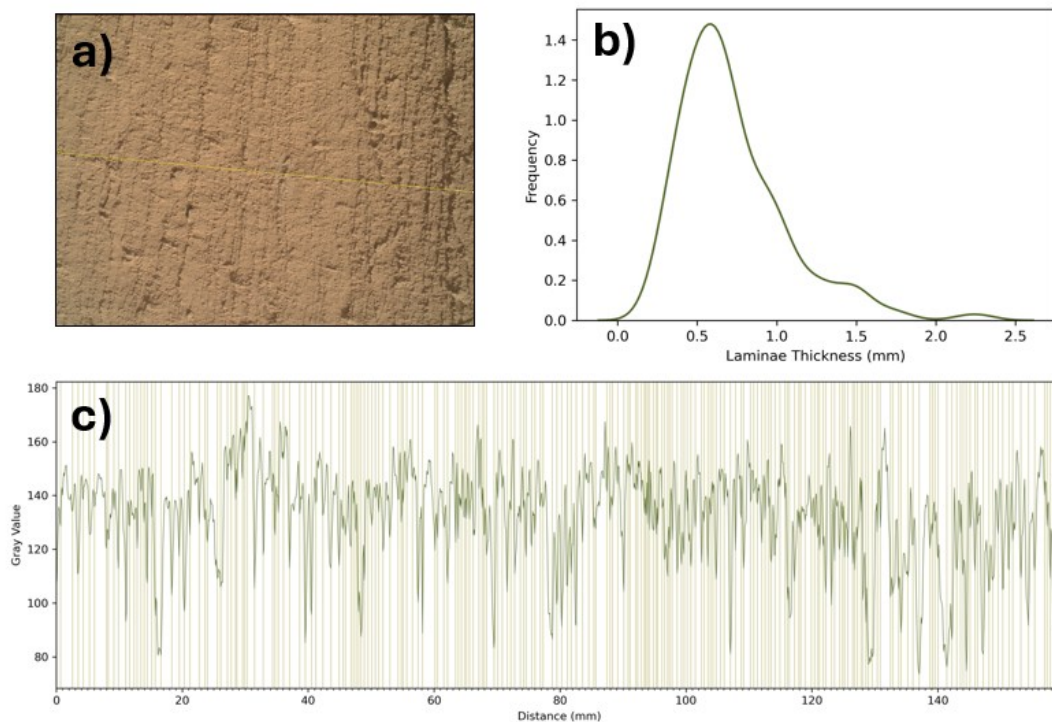


Figure I.4.1. Results of laminae thickness measurements for MAHLI image of target Marshall Falls, 5 cm stand off, taken on Sol 3596. (a) MAHLI image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

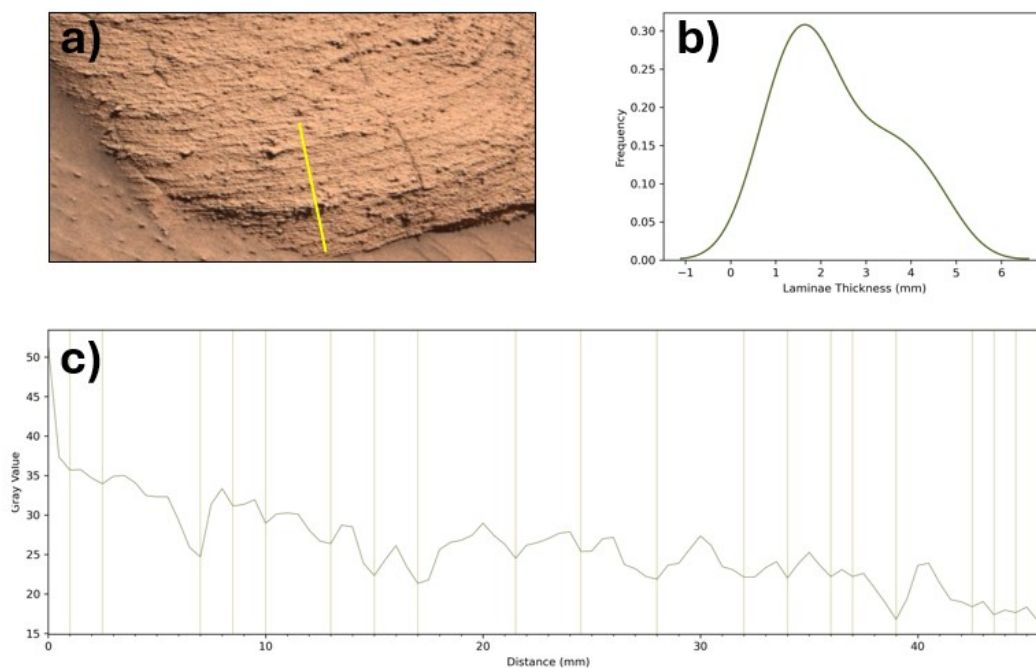


Figure I.4.2. Results of laminae thickness measurements for Mastcam image 103899. (a) Mastcam image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

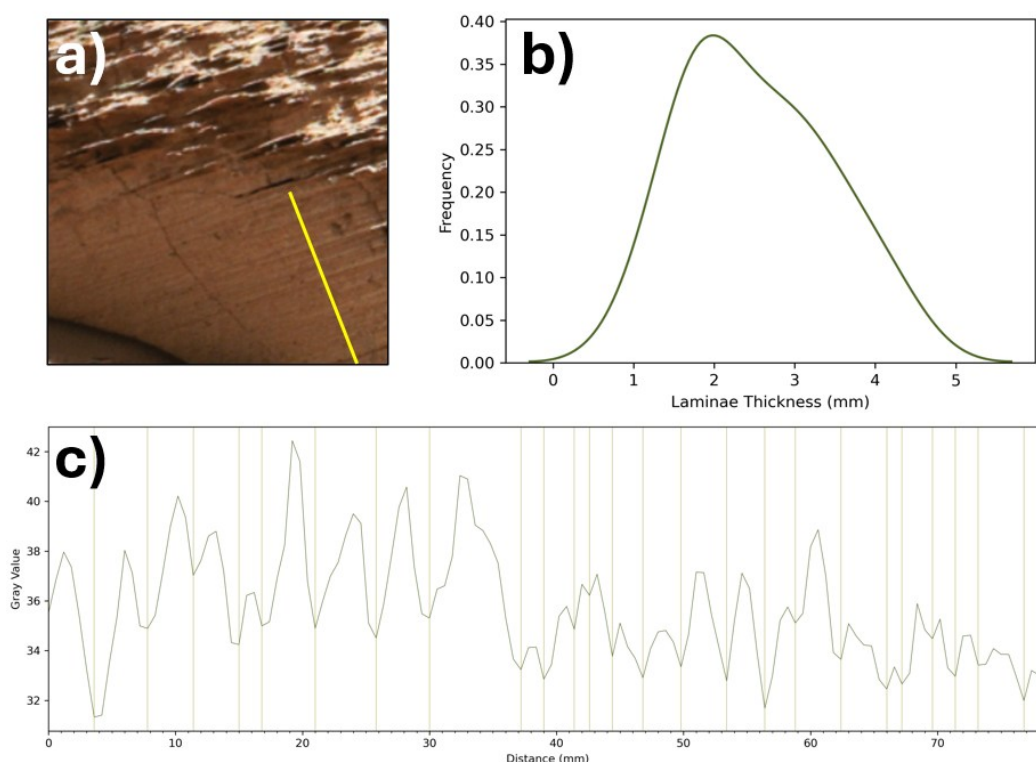


Figure I.4.3. Results of laminae thickness measurements for Mastcam image 103951. (a) Mastcam image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

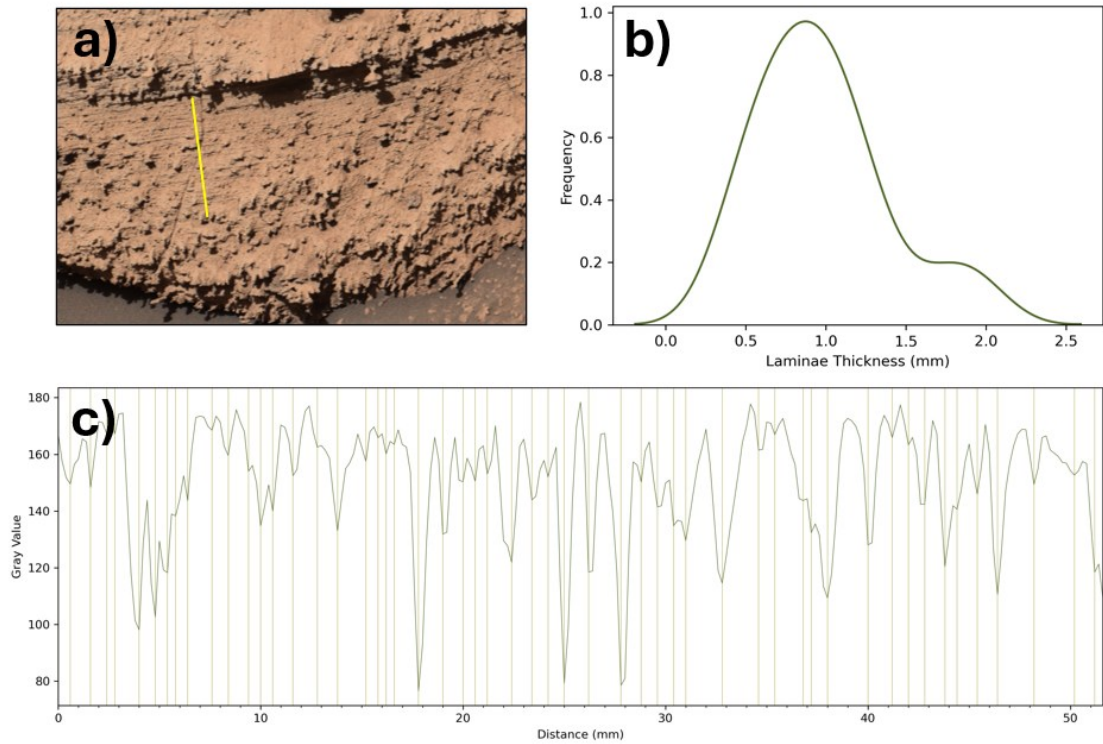


Figure I.4.4. Results of laminae thickness measurements for Mastcam image 104018. (a) Mastcam image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

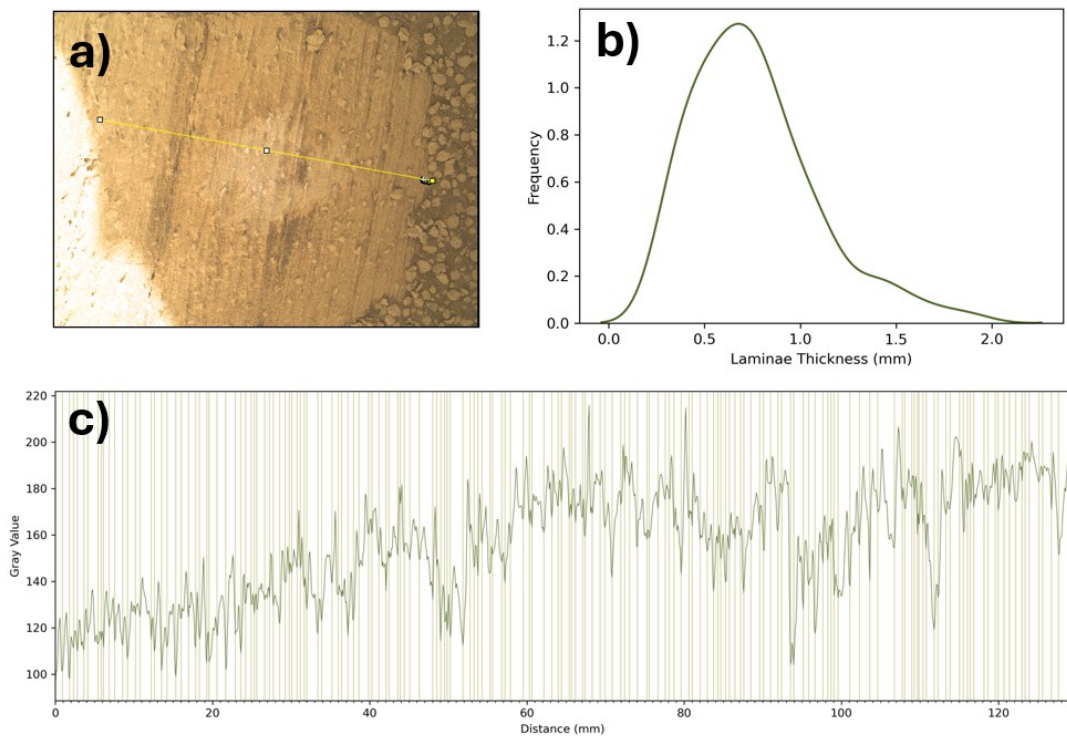


Figure I.4.5. Results of laminae thickness measurements for MAHLI image of Kastia Spring target, 5 cm stand off. (a) MAHLI image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of

peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

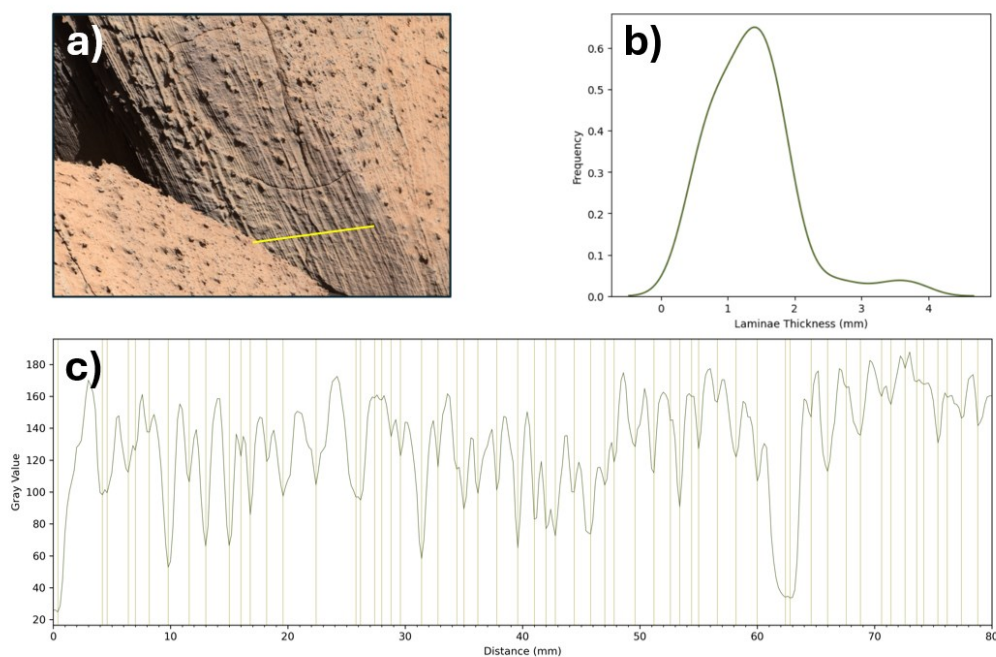


Figure I.4.6. Results of laminae thickness measurements for Mastcam image 104420. (a) Mastcam image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

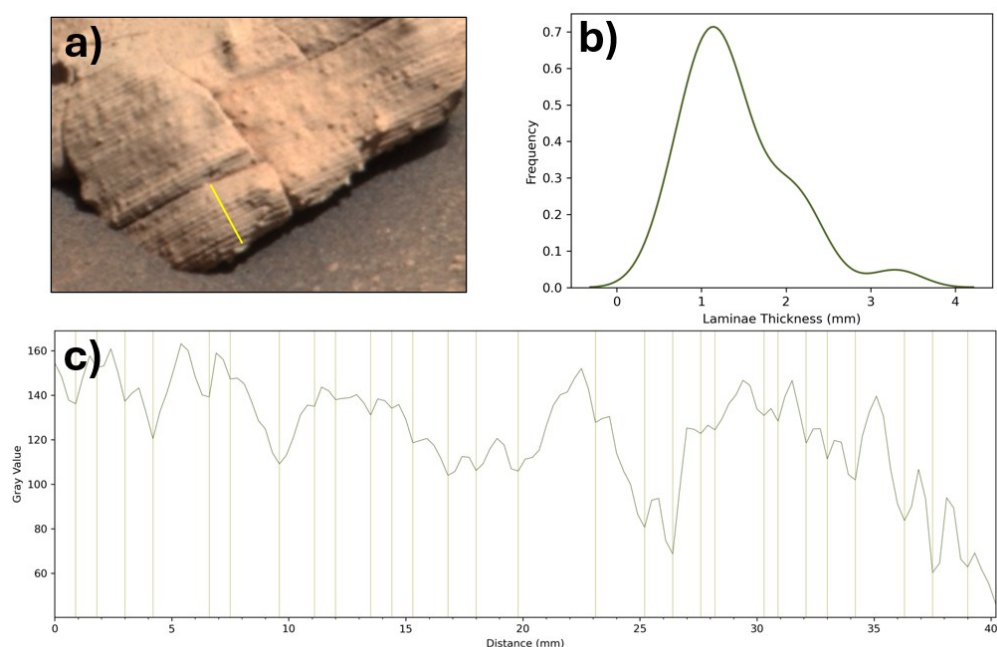


Figure I.4.7. Results of laminae thickness measurements for Mastcam image 104608. (a) Mastcam image with transect line used for laminae thickness measurements. (b) size distribution of laminae thicknesses. (c) Results of peak detection on grayscale pixel intensity using the `find_peaks` function in the Python `scipy` library (Virtanen et al., 2020), with peak spacing used as a proxy for laminae thickness. Image credits: NASA/JPL-Caltech/MSSS.

Appendix II. Tables

II.1 Complete table of missions to Mars

Mission	Launch Date	Country / Agency	Type(s)	Remarks / Outcome
Marsnik 1	10-Oct-60	USSR	Flyby	Launch failure - did not reach Earth orbit
Marsnik 2	14-Oct-60	USSR	Flyby	Launch failure
Sputnik 22	24-Oct-62	USSR	Flyby	Launch failure
Mars 1	01-Nov-62	USSR	Flyby	Communications lost before flyby
Sputnik 24	04-Nov-62	USSR	Lander attempt	Launch failure
Mariner 3	05-Nov-64	USA	Flyby	Launch failure (shroud didn't separate)
Mariner 4	28-Nov-64	USA	Flyby	Successful; first returned pictures of Mars
Zond 2	30-Nov-64	USSR	Flyby	Radio failure; minimal data returned
Mariner 6	25-Feb-69	USA	Flyby	Successful
Mariner 7	27-Mar-69	USA	Flyby	Successful
Mars 1969A	27-Mar-69	USSR	Orbiter	Launch failure
Mars 1969B	02-Apr-69	USSR	Orbiter	Launch failure
Mariner 8	09-May-71	USA	Orbiter	Launch failure
Cosmos 419	10-May-71	USSR	Orbiter	Launch failure
Mars 2	19-May-71	USSR	Orbiter + Lander	Partial success: orbiter succeeded, lander crashed
Mars 3	28-May-71	USSR	Orbiter + Lander	Lander briefly transmitted after landing
Mariner 9	30-May-71	USA	Orbiter	First spacecraft to orbit Mars; successful
Mars 4	21-Jul-73	USSR	Orbiter	Failed to enter stable orbit
Mars 5	25-Jul-73	USSR	Orbiter	Entered orbit; failed after ~9 days
Mars 6	05-Aug-73	USSR	Flyby + Lander	Lander lost contact during descent
Mars 7	09-Aug-73	USSR	Flyby + Lander	Lander missed target due to spacecraft failure
Viking 1	20-Aug-75	USA	Orbiter + Lander	Successful

Viking 2	09-Sep-75	USA	Orbiter + Lander	Successful
Phobos 1	07-Jul-88	USSR	Orbiter + Phobos Lander	Spacecraft failure before reaching Mars
Phobos 2	12-Jul-88	USSR	Orbiter + Phobos Lander	Lost contact before lander deployment; orbiter partially successful
Mars Observer	25-Sep-92	USA	Orbiter	Failure: lost before Mars arrival
Mars Global Surveyor	07-Nov-96	USA	Orbiter	Successful; long service
Mars 96	16-Nov-96	Russia	Orbiter + Penetrators	Launch failure
Mars Pathfinder / Sojourner	04-Dec-96	USA	Lander + Rover	Successful; first rover on Mars
Nozomi	03-Jul-98	Japan	Orbiter	Failed to enter Mars orbit
Mars Climate Orbiter	11-Dec-98	USA	Orbiter	Failed (navigation error)
Mars Polar Lander	03-Jan-99	USA	Lander	Lost contact during landing
2001 Mars Odyssey	07-Apr-01	USA	Orbiter	Successful; still operating
Mars Express	02-May-03	ESA	Orbiter	Successful
Beagle 2	02-May-03	UK / ESA	Lander	Contact lost; partial landing
Spirit (MER-A)	10-Jun-03	USA	Rover	Successful
Opportunity (MER- B)	08-Jul-03	USA	Rover	Successful
Mars Reconnaissance Orbiter	12-Aug-05	USA	Orbiter	Successful
Phoenix	04-Aug-07	USA	Lander	Successful
Fobos-Grunt / Yinghuo-1	08-Nov-11	Russia / China	Phobos Lander + Orbiter	Launch failure
Curiosity (Mars Science Laboratory)	26-Nov-11	USA	Rover	Successful; still operating
Mars Orbiter Mission (Mangalyaan)	05-Nov-13	India	Orbiter	Successful
MAVEN	18-Nov-13	USA	Orbiter	Successful
Trace Gas Orbiter (ExoMars)	14-Mar-16	ESA / Roscosmo s	Orbiter	Successful
Schiaparelli EDM	14-Mar-16	ESA	Lander	Crash / contact lost before landing
InSight	05-May-18	USA	Lander	Successful
Hope (Emirates Mars Mission)	19-Jul-20	UAE	Orbiter	Successful

Tianwen-1 / Zhurong	23-Jul-20	China	Orbiter + Lander + Rover	Successful
Perseverance (Mars 2020) + Ingenuity	30-Jul-20	USA	Rover + Helicopter	Successful; ongoing operations

II.2 Table of Mastcam mosaics used in this work

Figure	Target Name	Instrument	Camera sequence / Product code	Sol Taken	End-of-drive location
2.4, 3.1a	“Orinoco,” “Chenapau,” and “Kukenan”	M100	MR_mcam102554	3562	3650
2.5a	“Camopi”	M100	MR_mcam103951	3807	3805
2.5b	“Cariacou”	M100	MR_mcam104131	3845	3843
2.5c	“Bela Vista”	M100	MR_mcam103061	3648	3646
2.6, 2.8, 3.10a, 3.12a, 3.14a-b	“Chenapau”	M100	MR_mcam103485	3727	3724
2.9, 2.12	“Orinoco” and “Kukenan”	M100	MR_mcam102672	3578	3576
2.10	“Chenapau”	M100	MR_mcam102724	3589	3579
2.11	“Bela Vista”	M100	MR_mcam102968	3635	3633
2.14, 2.15, 2.16	“Marker Band valley ledge”	M100	MR_mcam103501	3730	3728
2.17, 3.11	“Gediz Vallis ridge”	M100	MR_mcam103090	3653	3651
3.3, 3.6a, 3.6b	“Orinoco”	M100	MR_mcam103144	3662	3658
3.4a		M100	MR_mcam102699	3582	3579
3.4b		M100	MR_mcam102748	3594	3592
3.4c		M100	MR_mcam102757	3596	3594
3.5		M100	MR_mcam102097	3485	3483
3.6c, 3.6d		M100	MR_mcam102214	3507	3506
3.7a		M100	MR_mcam103352	3705	3703
3.7c		M100	MR_mcam102798	3602	3601
3.7e		M100	MR_mcam102841	3610	3608
3.8a	“Orinoco”	M100	MR_mcam102754	3596	3595
3.8d		M100	MR_mcam103019	3642	3640
3.9a-c	“360”	M100	MR_mcam103228	3642	3640
3.9 d-e		M100	MR_mcam103019	3642	3640
3.10b		M100	MR_mcam103386	3710	3708
3.10c	“Amapa”	M100	MR_mcam103387	3711	3708
3.10d		M34	MR_mcam103736	3771	3771
3.12c		M100	MR_mcam103808	3783	3781
3.13		M100	MR_mcam103899	3798	3797
3.14c		M100	MR_mcam103913	3800	3799
3.15a		M100	MR_mcam103951	3807	3805

3.15b-c	M100	MR_mcam104018	3819	3815
3.15d	M100	MR_mcam103987	3812	3810
3.15e	M100	MR_mcam103987	3812	3810
3.15g	M100	MR_mcam104069	3831	3815
3.16a	M100	MR_mcam103913	3800	3799
3.16b	M100	MR_mcam104131	3845	3843
3.17a	M100	MR_mcam103485	3727	3724
3.17c	M100	MR_mcam104297	3871	3870
3.17d	M100	MR_mcam104379	3885	3880
3.18a	M100	MR_mcam104278	3867	3865
3.18b	M100	MR_mcam104420	3892	3891
3.18c	M100	MR_mcam104328	3876	3873
3.18d	M100	MR_mcam104461	3898	3897
3.18e	M100	MR_mcam104403	3889	3887

II.3 Table of LD RMI mosaics used in this work.

Figure	Target Name	Camera sequence / Product code	Sol Taken	End-of-drive location
2.7a-c	“Orinoco scour”	ccam_05671	3671	3667
2.13	“Chenapau scour”	ccam_04582	3582	3579
3.8b	“Orinoco”	ccam_03625	3623	3608
3.12b	“Chenapau”	ccam_02728	3728	3727

II.4 Table of MAHLI mosaics used in this work.

Figure	Target Name	Camera sequence/Product code	Sol Taken	End-of-drive location
3.7b	Tapirapeco	3605MH0008410011300992C00_MhliImage_0717532103-09606-1	3605	3603
3.8e	Mamupi	3642MH0003690001301560R00_MhliZstack_0720811892-06470-1	3642	3640
3.9c	Mixiguana	3644MH0008550011301571C00_MhliImage_0720987511-47479-1	3644	3642
3.10e	Santa Elena de Uairen	3773MH0008410021304346C00_MhliImage_0732453957-35173-1	3773	3771
3.15f	Anortosito Repartimento	3810MH0001900001400097C00_MhliImage_0735732890-09593-1	3810	3808
2.5d, 3.16d, I.3, I.4.5	Kastria Spring	3866MH0001900011400878C00_MhliImage_0740708055-51665-1	3866	3865
I.4.1	Marshall Falls	3596MH0001900011300680C00_MhliImage_0716740069-45544-1	3596	3594

II.5 Table of Scour Measurements Additional Information

Scour	Way of measuring	Mastcam/Navcam image used	M100 spatial sampling (mm/pixel)	M34 spatial sampling (mm/pixel)
Chenapau Scour One	3D mesh	sol03727_MR_103485_chenapau_stereo_13x8_L0_R0_P_F	6.6±0.5	20±1
Chenapau Scour Two	3D mesh	sol03750_ML_103639_chenapau_stereo_7x1_L0_R0_PF	5.8±0.3	17±1
Orinoco Scour Two	Correlating MastCam images and HiRISE images	Sol03578_MR_102672_orinoco_kukenan_stereo_10x5_L0_R0_PF	19±1	56±4
Canta Detour Scour	3D mesh	sol03653_MR_103090_gediz_vallis_ridge_18x4_L0_R0_PF	3.6±0.2	11±1
Bela Vista Scour	3D mesh	sol03635_ML_102968_marcaca_stereo_113x1_L0_R0_PF.png	6.9±0.3	20±1
Orinoco Scour One	Correlating MastCam images and HiRISE images	Sol03578_MR_102672_orinoco_kukenan_stereo_10x5_L0_R0_PF	17±1	50±1
Chenapau Scour Three	Correlating MastCam images and HiRISE images	Sol03589_MR_102724_Chenapau_stereo_10x3_L0_PF	16±1	417±1
Exhumed Scour One	3D mesh	sol03730_MR_103501_scour_stereo_15x2_L0_R0_PM.png	6.5±0.5	20±1
Exhumed Scour Two	3D mesh	sol03730_MR_103501_scour_stereo_15x2_L0_R0_PM.png	7.4±0.3	22±1
Exhumed Scour Three	3D mesh & HiRISE images	sol03730_MR_103501_scour_stereo_15x2_L0_R0_PM.png	13±0	39±1

Note: Spatial sampling was determined based on the distance between the scours and the rover, following the method described by Bell et al. (2017). For scours located within 50 metres of the rover, distances were measured using a 3D Navcam mesh (Alexander et al., 2006). For scours beyond 50 metres, measurements were obtained using ArcGIS Pro 3.1.0. The uncertainty in spatial sampling accounts for the range between the maximum and minimum distances measured from the scours to the rover, relative to the average distance.

II.6. Table showing the summary of stratigraphic unit elevations across major buttes and along the rover traverse within Marker Band valley.

Unit	Elevations in Orinoco	Elevations along the traverse	Sols along the traverse	Elevations in Chenapau	Elevations in Bolivar	Elevations in Deepdale	Elevations in Owenteik
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Unit A	-3870 to -3856	to -	-3877 to -3865	to -	3560 to -3607	N/A	-3894 to -3867	to -	-3898 to -3860	N/A
Unit B	-3856 to -3852	to -	-3864 to -3859;	to -	3635 to -3642;	N/A	-3867 to -3862	to -	-3860 to -3855	N/A
			-3863 to -3860	to -	3687 to -3708					
Unit C	-3852 to -3852	to -	-3860		3641	N/A	-3862 to -3862	to -	N/A	N/A
Unit D	-3852 to -3851	to -	-3859; -3858 to -3852	-	3638 to -3645;	-3857 to -3854	-3861 to -3862	to -	-3856 to -3857	N/A
					3667 to -3672					
Unit E	-3851 to -3845	to -	-3859 to -3852	to -	3645 to -3667;	-3854 to -3849	-3861 to -3854	to -	-3856 to -3847	-3851 to -3847
					3771 to -3776					
Unit F	-3845 to -3840	to -	-3848 to -3838	to -	3776 to -3796	-3849 to -3844	N/A		N/A	-3847 to -3842
Unit G	-3840 to -3834	to -	-3838 to -3832	to -	3796 to -3805	-3844 to -3835	N/A		N/A	-3842 to -3835
Unit H	-3834 to -3828	to -	-3832 to -3826	to -	3801 to -3817*	-3835 to -3829	N/A		N/A	-3835 to -3834
Unit I	-3829 to -3821	to -	-3826 to -3812	to -	3812* to -3867	-3830 to -3822	N/A		N/A	N/A
Unit J	-3821 to -3807	to -	-3812 to -3798	to -	3867 to -3901	-3822 to -3807	N/A		N/A	N/A
Unit K	-3807 to -3798	to -	-3796 to -3773	to -	3901 to -3944	-3807 to -3794	N/A		N/A	N/A

*Travelled along approximate unit contact

Appendix III. Software and Techniques

III.1 Construction of DTMs and Orthomosaics

The $\sim 1/\text{m}$ per pixel HiRISE DEMs were created from suitable HiRISE stereo pairs taken with an emission angle difference of $>5^\circ$ under similar lighting conditions, calibrated to the ($>400 \text{ m/pixel}$) MOLA DEMs following the method described in Kirk et al. (2008). First, the public-domain Integrated Software for Images and Spectrometers (ISIS3) software was used to pre-process stereo pairs to radiometrically calibrate the images and provide supporting SPICE (Spacecraft and Planetary ephemerides, Instrument C-matrix and Event) files with important data for the image pixels, including latitude, longitude, emission angle, and local solar time. Then, the processed images and supporting information were imported into the BAE Systems photogrammetric software package SOCET SET (v5.6.0; <http://www.geospatialexploitationproducts.com/content/socet-set-v56/>) where matching pixels of the two stereopairs were tied together, and vertical ground control information is added from MOLA gridded data and MOLA track points (if the points cross the images). A DEM was then created by extracting the individual pixel elevations from the images at a specific resolution through the algorithm NGATE. After the initial DEM was created, a second algorithm (ATE) was then run, which reduced the noise in the DEM. Orthorectified image products were then generated by removing the distortion in the images, taking into account the shift in pixel location. Finally, the orthorectified images and the DEM were exported back into ISIS, where they were converted into cube files, readable in a GIS.

III.2 Construction of Viewsheds

Viewsheds were constructed using the Viewshed tool in *ArcGIS Pro 3.1.1* using the HiRISE DEM (Calef and Parker, 2016), rover localisations, rover mast height, and rover elevation. The observer height was set to approximate the Mastcam viewing position above the ground. Where the viewshed was intended to represent a specific image rather than a full panorama, the output was clipped with reference to specified Mastcam field of views.

III.3 White Colour Balancing

A white colour balance was performed using *Adobe Photoshop* to ensure colour consistency between images, to highlight subtle contrasts that record changes in surface properties, and to provide Earth-like lighting conditions for better representation of colours for interpretation and comparisons with terrestrial analogues.

III.4 Measuring Dip and Strike

Apparent directions and azimuths of cross-bedding were estimated using stereo ranging and bearings of points selected on Navcam or Mastcam mosaics, plus the Cartesian coordinates of the point relative to the rover, allowing the point to be projected onto HiRISE images. Planetary Robotics 3D viewer

(PRo3D) (Paar et al., 2017) was used to visualize stereo-images collected by the Navcam and Mastcam instruments. Ordered Point Clouds were generated from processed Navcam and Mastcam frames, which were viewed and analysed in PRo3D (Paar et al., 2017). However, the absence of three-dimensional exposures of cross-bedding in Marker Band valley meant that only the apparent dip could be calculated, and dips of cross-bedding described in this thesis should not be considered representation of the true dip of the cross-bedding.

III.5 Data Availability

The colour HiRISE mosaic of Gale crater used in Chapters 2 and 3 is available through Calef and Parker (2016). Additional data from HiRISE (McEwen et al., 2007) and CTX (Malin et al., 2007) images can be accessed at the NASA Planetary Data System Geosciences Node (<https://pds-geosciences.wustl.edu/missions/mro/default.htm>). All Mastcam (Malin et al., 2017) and MAHLI (Edgett et al., 2012) images referenced in this thesis can be found at the NASA Planetary Data System Cartography and Imaging Sciences Node (<https://pds-imaging.jpl.nasa.gov/volumes/msl.html>).